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Recherches sur l'ozone atmosphérique et la température de la stratosphère en Laponie Suédoise

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Abstract

Measurements of atmospheric ozone have been made in Abisko (Swedish Lappland) during two different periods, one extending from December 1934 to March 1935, the other beginning with August 1949. Measurements of "mean temperature" have shown an annual temperature variation the amplitude of which is in excess of 100° C and which is due to the prolonged insolation of the stratosphere in high latitudes. It follows from the discussion of the results that it is necessary to determine the annual variations of the vertical distribution of ozone and of temperature in the stratosphere of those regions.

On sait depuis longtemps que les latitudes élevées présentent un intérêt particulier pour certaines études de géophysique, et avec le développement de la science météorologique cet intérêt s'est étendu à toutes les études sur l'atmosphère.

L'ozone atmosphérique est un problème qui touche par certains de ses aspects à la météorologie, à qui il peut fournir de précieux renseignements; et par d'autres à la géophysique et aux relations entre la Terre et le Soleil; aussi n'est-il pas surprenant que des chercheurs français soient venus travailler en Laponie Suédoise pour compléter notre connaissance de l'ozone à laquelle Ch. Fabry et ses élèves ont apporté une si large contribution. Nous parlerons ici des séjours effectués à Abisko pendant l'hiver 1934—35, par MM. Barbier, Chalonge, Dauvillier et Vassy, et par nous-mêmes pendant l'été 1949.

1. Position du problème; anciennes recherches

L'origine de ces recherches remonte à 1932. Faisant partie de l'organisation de l'Année Polaire Internationale de 1932, une mission française s'était établie au Scoresby Sund (Groënland). Dauvillier, qui était dans ce groupe, avait accompli des mesures régulières de la teneur en ozone de l'air au voisinage du sol, mesures effectuées par voie chimique, et ses dosages révélèrent à partir du 1er décembre une teneur de plus en plus forte qui dépassa 10 fois la valeur moyenne, pour revenir le 1er mars à des teneurs normales (DAUVILLIER 1934) Ces fortes concentrations pendant la nuit polaire lui firent attribuer l'origine de l'ozone atmosphérique aux électrons auroraux, et c'est en partie en vue de vérifier cette hypothèse que fut entreprise la mission 1934—1935 à Abisko qui séjourna à la »Naturvetenskapliga Station» grâce à l'aimable intervention du Prof. Siegbahn.

Les mesures se partageaient en deux séries: mesure de la teneur en ozone des couches basses, et mesures de l'»épaisseur réduite» de l'ozone de toute l'atmosphère; les 2 séries de mesures étaient faites par voie spectrographique avec le même instrument, les premières sur une source terrestre à courte distance, un tube à hydrogène placé à la Touriststation (1 500

mètres environ), les secondes sur des étoiles, de préférence de type A ou B, dont le spectre ultraviolet est plus simple et avait été étudié auparavant par Barbier, Chalonge et E. Vassy. Disons que les mesures d'épaisseur réduite continuaient les importants travaux de Dobson sur les variations de l'ozone en fonction de la latitude et de la saison. Concurremment Dauvillier effectuait des dosages par voie chimique de l'ozone des couches basses.

Les mesures spectrographiques de la teneur en ozone de l'air au niveau du sol effectuées chaque jour du 29 décembre 1934 au 6 mars 1935 ont donné les résultats suivants (Barbier, Chalonge, Vassy, 1936): la valeur moyenne, exprimée suivant l'usage en microns par km (épaisseur d'ozone pur à la pression normale qui produirait la même absorption que l'ozone de l'air observé pour un trajet de 1 km), était de 19,7 μ par km; les variations autour de cette moyenne étaient peu remarquables, parfaitement incohérentes, les valeurs extrêmes étant 30 μ le 4 janvier, et 7 μ le 24 février.

La valeur moyenne était du même ordre de grandeur que celle observée aux latitudes moyennes à Lauterbrunnen (Suisse), et inférieure à celles observées à plus haute altitude, Jungfraujoch et Arosa (Chalonge, Vassy, 1934 a).

Quant aux valeurs individuelles, aucune corrélation n'a pu être trouvée, ni avec divers éléments météorologiques, ni avec l'épaisseur totale d'ozone, ni avec la nature des masses d'air. On n'avait pas observé de variation saisonnière ou d'influence de l'activité aurorale.

Les résultats de Dauvillier au Scoresby Sund ne se retrouvaient donc pas à Abisko, ainsi qu'il ressortait également de ses propres mesures chimiques (1935).

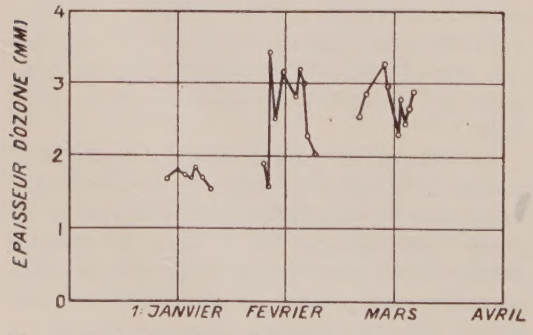

Fig. 1. Epaisseur réduite de l'ozone à Abisko (décembre 1934—mars 1935).

Les mesures d'épaisseur totale (Barbier, Chalonge, Vassy, 1935 a) sont représentées par le graphique de la fig. 1; les intervalles sans mesures sont dus à la présence de la Lune qui empêchait les observations, le spectrographe étant d'un modèle à prisme-objectif mis au point pour ces recherches (Chalonge, Vassy, 1934 b) et dont l'absence de fente ne permet pas les mesures en présence de lumières parasites. Le graphique montre la croissance de l'ensemble à l'approche du printemps, en liaison avec la variation saisonnière. Quant aux variations d'un jour à l'autre, elles se sont révélées en liaison avec les grands mouvements du front polaire ainsi que les premiers travaux de Dobson l'avaient déjà montré (1930). D'une façon générale, les variations journalières sont liées aux phénomènes météorologiques; elles se superposent aux grandes variations d'origine géophysique que l'on fait apparaître par la considération de moyennes, mensuelles par exemple.

2. Utilisation des données de l'observation

Les résultats obtenus en 1934—35 auraient donc été assez décevants si les recherches en étaient restées là. Mais en 1936, à la suite de mesures effectuées au laboratoire sur les coefficients d'absorption de l'ozone à differentes températures, l'un de nous (Vassy, 1936) a montré que, par suite de la variation linéaire des coefficients correspondants aux minima d'absorption, il était possible de définir une température moyenne de l'ozone atmosphérique (ozone de toute l'atmosphère), et il a donné également la méthode permettant de déterminer cette température moyenne à partir du spectre d'absorption (Vassy, 1937); alors que jusque là on avait dû se borner à montrer que l'ozone atmosphérique était à une température inférieure à 18° C (Barbier, Chalonge, Vassy, 1935 b), on pouvait maintenant donner la température moyenne de l'ozone sous nos latitudes, soit — 30° C.

En possession de cette méthode, nous en avons fait l'application systématique au cours d'une mission au Maroc en 1937; et celà nous permit de montrer que cette donnée, s'ajoutant à l'épaisseur réduite de l'ozone, peut servir à déterminer la nature d'une masse d'air, et, dans certains cas, peut faciliter la prévision météorologique (A. et E. Vassy, 1938).

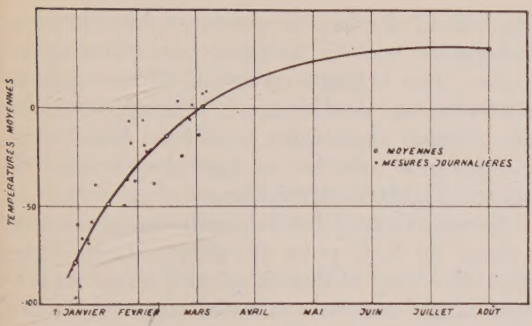

Fig. 2. Variation annuelle de la température moyenne de l'ozone à Abisko.

Barbier et Chalonge (1939) appliquèrent la méthode ensuite à un ensemble de mesures groupant diverses stations d'observations. Ayant extrait de leur travail les résultats concernant Abisko, que la fig. 2 représente, nous avons alors trouvé (A. et E. Vassy, 1939) une importante variation saisonnière de la température moyenne à la latitude d'Abisko. On constate en effet un relèvement systématique et important de la température moyenne en fonction du temps entre le 27 décembre 1934 et le 6 mars 1935. Or, pour le 27 décembre, à l'altitude de 25 km qui correspond à peu près au centre de gravité de la couche d'ozone, la durée du jour n'était que de 4 heures $^1/_2$, alors que le 6 mars elle atteignait 12 heures. Malgré les fluctuations dues à la circulation atmosphérique, cette variation montre donc une influence très nette du rayonnement solaire sur la température moyenne de l'ozone, et par suite sur la température de la stratosphère puisqu'en raison des pressions existant au-dessous de 50 km, l'ozone ne peut avoir une température différente des autres constituants de l'atmosphère.

Pour chiffrer ce relèvement de température, nous avons fait les moyennes mensuelles des températures moyennes de l'ozone atmosphérique; elles sont les suivantes:

fin décembre	— 79° C
janvier	— 49° C
février	— 14,5° C
début mars	+ 1° C

3. Recherches récentes

Ayant ainsi trouvé sur la température moyenne de l'ozone une importante influence du Soleil qui semble particulière aux régions arctiques, il nous a paru intéressant de voir ce que devenait cette température moyenne durant les mois d'été où la stratosphère des hautes latitudes reçoit le rayonnement solaire pendant une durée prolongée et même sans interruption pendant plusieurs semaines à l'époque du solstice.

1. Résultats

Aussi avons-nous profité d'un séjour à l'Institutet för högspänningsforskning d'Uppsala, dirigé par le Prof. Norinder, séjour effectué sur les fonds du Conseil des Sciences Naturelles de l'Etat de Suède, pour déterminer par la même méthode la température moyenne de l'ozone en été. Des mesures ont été faites à Uppsala en août 1949, et nous avons séjourné 4 jours à Abisko où, sur l'initiative du Prof. Tiselius, nous avons reçu l'hospitalité la plus agréable à la Naturvetenskapliga Station, du 31 juillet au 4 août 1949.

Malheureusement, en raison du mauvais temps qui a régné à peu près constamment pendant notre séjour à Abisko et aussi à Uppsala, les spectres d'Abisko sont pris à travers une atmosphère chargée d'humidité, et nous avons déjà attiré l'attention sur les bandes d'absorption de la vapeur d'eau dans le proche ultraviolet qui pourraient fausser les mesures surtout si l'on opère à basse altitude.

Néanmoins nous avons pu obtenir les valeurs cherchées (A. et E. Vassy, 1950): à Uppsala, le 6 août l'épaisseur réduite était 2,72 mm et la température moyenne + 25° C; à Abisko l'épaisseur réduite était le 1er août 2,74 mm et la température + 30° C.

Cette valeur de + 30° C se place de façon très correcte sur la courbe de température obtenue en 1934 et extrapolée jusqu'en août (fig. 2) confirmant ainsi l'influence considérable du Soleil sur la température de la stratosphère et due sans doute à l'éclairement ininterrompu.

Il est évidemment dommage que nous disposions de si peu de valeurs pour ce mois d'août. Nous pouvons cependant accorder une certaine confiance à cette valeur de + 30° C car la situation météorologique était stable. C'est ce qui résulte de l'analyse de la situation aérologique que MM. Ångström et Björkdal, Directeur et Directeur-Adjoint de l'Institut météorologique suédois, ont bien voulu nous communiquer et nous les en remercions très vivement. Du 30 juillet au 2 août les masses

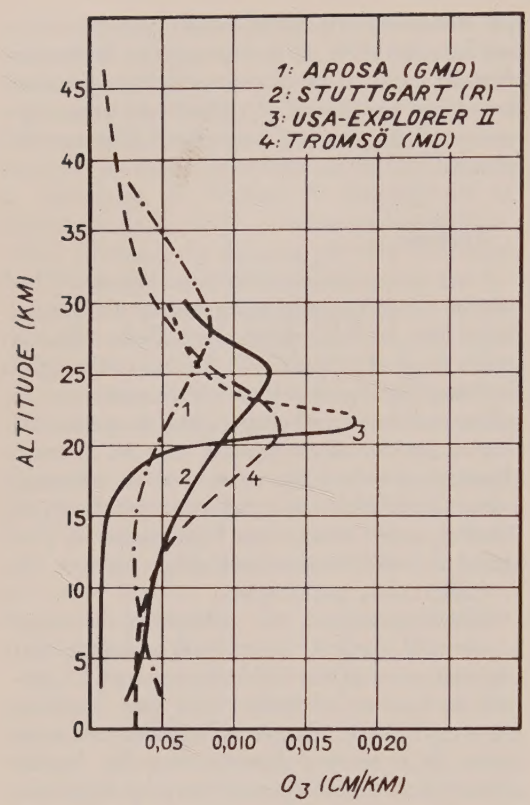

Fig. 3. Distribution verticale de l'ozone à Tromsø (Meetham et Dobson).

d'air étaient à peu près homogènes, aucun front ne traversant la région. Si l'origine des masses d'air ne peut être donnée avec certitude, il est probable qu'il s'agit d'air polaire maritime âgé. Pas de transformation par subsidence. La température donnée nous parait donc correspondre assez bien à la saison.

Une variation annuelle de température aussi importante est très remarquable; elle ne doit pas être sans influer sur les mouvements à grande échelle de l'atmosphère. Aussi serait-il bon de préciser à quelles altitudes elle se rapporte.

2. Interprétation

C'est pourquoi il est intéressant de comparer ces valeurs données par l'ozone avec celles obtenues par Rolf au moyen de sondages (1932). En effet, les températures directement accessibles par les sondages se limitent aux couches inférieures à l'altitude maximum atteinte par les ballons, qui varie entre 15 et 20 km. L'ozone qui s'étend plus haut nous permet donc de nous renseigner sur les couches plus élevées, et même, dans la limite où nous avons sa répartition verticale, de déterminer la température de ces couches supérieures à 30 km. Mais cette comparaison avec les sondages devait nous réserver d'intéressantes surprises.

Si nous considérons la courbe moyenne des valeurs de Rolf pour décembre, la température de — 79° C donnée pour l'ozone est en bon accord avec cette courbe supposée extrapolée au-dessus de 20 km; si cette extrapolation est correcte, on atteindrait — 79° C à l'altitude de 30 km environ; nous avions trouvé de même qu'à la latitude de 45°, la température moyenne est celle que l'on obtient en extrapolant les sondages jusqu'à 30 km (A. et E. Vassy, 1941).

Mais pour les autres mois, le résultat est bien différent. En janvier, la courbe de température est continuellement décroissante; à la température — 49° C correspond la seule altitude de 7,5 km; en février de même, la température — 14,5° correspond à 2 km. Ces résultats sont évidemment absurdes, la plus grande partie de l'ozone se trouvant certainement, quelle que soit sa répartition, à des altitudes supérieures à 15 km. En mars, le gradient de température est très légèrement croissant entre 12 et 20 km; mais pour trouver en extrapolant, dans l'hypothèse d'un gradient uniforme, une température de + 1° C, il faudrait atteindre environ 100 km; en août il en serait à peu près de même. Ce résultat n'est donc pas acceptable non plus, les altitudes de 100 km ne pouvant contribuer, s'il y a de l'ozone, que pour une part négligeable à la température moyenne en raison de la très faible pression.

En conservant au-dessous de 18 km les températures données par les sondages, pour retrouver les températures élevées indiquées par l'ozone atmosphérique, il est indispensable de faire deux hypothèses simultanées:

1°) au-dessus de 18—20 km la température de l'atmosphère se relève rapidement;

2°) la plus grande partie de l'ozone se trouve au-dessus de 20 km.

Il semble donc que la répartition (voir fig. 3) donnée par Meetham et Dobson (1935) ne convienne pas, du moins pour les journées où nous avons fait nos mesures d'ozone. Par contre, aux latitudes moyennes en utilisant la répartition donnée pour Arosa et en extrapolant les sondages de cette latitude, on peut calculer une

température moyenne qui est, à 5° près, la température moyenne de l'ozone atmosphérique, — 30° C (VASSY, 1936).

Les régions de haute latitude posent donc un double problème: celui de la variation de la température avec l'altitude; celui de la variation de la répartition verticale de l'ozone avec la latitude et peut-être aussi avec la situation météorologique.

Il se pose encore un autre problème; au cours de notre étude sur l'origine de l'ozone atmosphérique (A. et E. VASSY, 1941) où nous avions montré que le facteur température jouait un rôle important, nous avions noté le cas anormal des régions polaires pour les mois d'été. Nous avions alors supposé que l'ozone se trouvait plus bas, ainsi que le montrait la répartition donnée par Meetham et Dobson, ce qui lui aurait donné une température moyenne inférieure et par suite faisait revenir les points figuratifs sur la droite. Or, nos dernières mesures nous montrent au contraire une température moyenne élevée. Il y a donc là encore un point à éclaircir.

Pour terminer, il nous est particulièrement agréable de remercier les organismes suédois qui, dans un large esprit de coopération scientifique internationale, nous ont rendu possible ces recherches, et les nombreuses personnes dont l'aide amicale et dévouée a grandement facilité notre travail.

Paris, le 11 mars 1950.

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Discussion of the Origin of the Terrestrial and Solar Magnetic Fields

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(Manuscript received 21 January 1950).

Abstract

The origin of the terrestrial and solar magnetic fields is discussed. An original meridional field can by hydrodynamic motion get an east-west component. It is pointed out that this component cannot exceed the order of magnitude of the original field. If it does, the field becomes unstable and an increase of the meridional component is caused. The possibility of finding an amplifying mechanism of this type is considered.

1. The origin of the terrestrial and solar magnetic fields is still one of the most puzzling problems of cosmical physics. Of the many different hypotheses that have been proposed, LARMOR's idea that the fields are due to internal motions seems to be the most attractive one [1]. Recent work has shown that a magnetic field could be amplified very much by hydrodynamic motions [2] [3] [4]. For example, it seems possible to account for the strong magnetic fields of sunspot by assuming that motions in the interior of the sun [2] have amplified the rather weak general solar field. In a similar way it is likely that in interstellar space an originally weak field has been amplified by hydrodynamic motions [4]. In analogy with these processes it seems reasonable that it should be possible to account even for the terrestrial and solar magnetic fields in terms of some sort of magneto-hydrodynamic process.

2. In the discussion of the general magnetic fields a paper by COWLING has attracted much interest [5]. COWLING draws the attention to the fact that the main problem is how new magnetic lines of force are created. A certain line of force may be deformed to any configuration by hydrodynamic motion and in this way a given magnetic field may be amplified. At the same time, however, the structure of the field becomes more and more complicated. The only way to amplify a field without changing its general character is to breed new lines of force. Under the conditions in the interior of the earth and sun COWLING is inclined to consider such a breeding as impossible.

Extensive investigations have been carried out by ELSASSER and BULLARD [6]. They doubt that COWLING's conclusion is valid without restrictions and try to find a rather complicated system of whirls by which an amplification should be possible. Their attempts have had no definite success so far. Upon the whole one has the impression that a phenomenon of such a general character as the terrestrial and solar fields should be due to a mechanism more simple than those proposed by these authors.

The kinematics of a field amplification

3. In order to find how the general magnetic fields of the earth and the sun are generated and maintained, it is advisable to start by trying to find a cyclic process by which a magnetic field could be amplified.

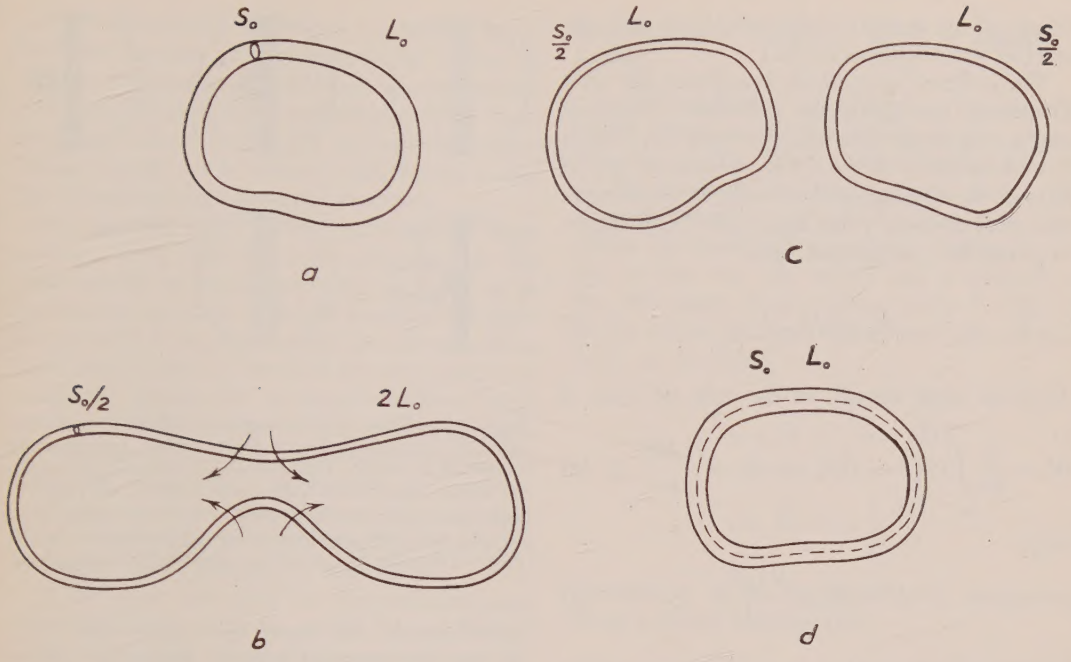


Fig. 1. The magnetic field in a closed tube of force can be amplified by hydrodynamic motion in the following way. The initial tube (*a*) is drawn out to the double length (*b*) and by a motion of the type marked by arrows this tube is split up into two tubes (*c*). These are placed side by side thus forming a tube (*d*) of the same shape as the initial tube but with the magnetic field doubled.

. It is rather easy to outline a kinematically possible process but to find the dynamics leading to this process is much more difficult. Consider a closed infinitely thin tube of force in an incompressible and infinitely conducting liquid (fig. 1 a). The circumference of the lines of force be L_0 and the cross-section of the tube S_0. To simplify the discussion we suppose that S_0 is constant. Then the volume of the tube of force is $V_0 = L_0 \cdot S_0$. As the liquid has an infinite conductivity no motion perpendicular to the lines of force is possible, and as the liquid is incompressible the volume U must remain constant. If the magnetic field is H_0, the flux is $\Phi_0 = S_0 H_0$.

We want to find a process which ends in a tube of the same dimensions but with an increased magnetic field. This could be accomplished in the following way. Let hydrodynamic motions lengthen the tube to $2L_0$. As V_0 and Φ_0 remain constant, this means that the cross-section diminishes to $S_0/2$ and the magnetic field increases to $2H_0$. By a motion of the type shown in fig. 1 b it is possible to split up this ring into two rings, each with the length L_0 (fig. 1 c). These two rings are placed side by side, thus forming one ring with the length L_0 and cross-section S_0, but with the magnetic field $2H_0$ (fig. 1 d). This process could be repeated indefinitely thus leading to an unlimited amplification of H.

4. The process considered refers to lines of force which are closed in the liquid, but it is also possible to increase the magnetic field by a process starting by open lines of force. In order to show this let us suppose that in an infinitely conducting sphere of incompressible liquid there is a uniform tube of flux from one pole to the other (fig. 2 a). If the upper half of this is rotated in relation to the lower half the magnetic lines of force will be twisted, so that the axial field H_0 is superimposed by a tangential field H_t. When the twisting exceeds a certain limit, the tube of force tends to make a loop (Fig. 2 b). The mechanical analogy of this phenomenon is well-known: if a wire is twisted, a series of loops are formed. (This is easily demonstrated by rotating the micro-

phone of an ordinary table telephone and observing the telephone cord.)

The order of magnitude conditions for loop formation can easily be calculated. Suppose that a cylindrical tube of force with the length L and radius $= R$ ($\ll L$) is twisted in one of its ends by an angle φ. In a volume element at the axial distance r the axial field H_a is superimposed by a tangential field

$$H_t = H_a \frac{r\varphi}{L} \tag{1}$$

Thus the total energy of the tube of force is

$$W_a = \frac{L}{8\pi}\int_0^R (H_a^2 + H_t^2)\, 2\pi r\, dr = \frac{H_a^2 R^2}{8}\cdot Y \tag{2}$$

with

$$Y = L + \frac{R^2\varphi^2}{2L} \tag{3}$$

If the tube of force changes its shape by making a loop, its length increases by at least $\Delta L = 2\pi R$. At the same time the twist changes by $\Delta\varphi = -2\pi$. As the volume $\pi R^2 L$ remains unchanged the radius changes by $\Delta R = -\frac{R}{2L}\Delta L = \frac{\pi R^2}{L}$. Further the flux $\pi R^2 H_a$ also remains constant, which means that the magnetic field changes by $\Delta H_a = \frac{2\pi H_a R}{L}$.

The condition that the total energy should not change gives

$$\Delta W_a = 0 \tag{4}$$

or

$$\varphi = (\sqrt{5} - 1)\, L/R \tag{5}$$

When φ exceeds this value the conditions for loop formation are favourable. By introducing (5) into (1) this approximate calculation shows that as soon as the tangential field exceeds the order of magnitude of the axial field a loop will be formed. Unlike the conditions in a telephone cord this loop will separate because the lines of force of the initial string will close again and so will the loop itself (fig. 2 c).

Let hydrodynamic motion draw out part of this loop to the surface (fig. 2 d), where

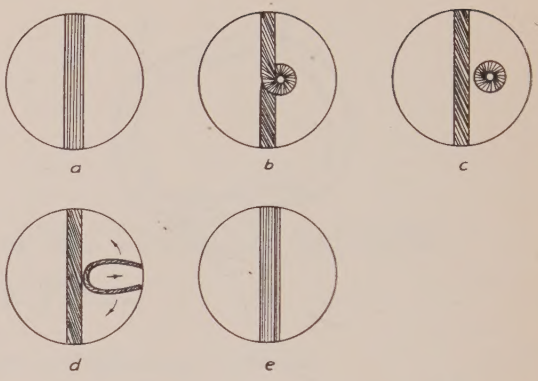


Fig. 2. Amplification of a magnetic field by loop formation. A magnetic tube of force within a conducting sphere (*a*) is twisted. This leads to the formation of a loop (*b*) which separates from the initial tube (*c*). By a hydrodynamic motion the loop may be split up and brought side by side with the initial tube (*d*) thus restoring the shape of the tube (*e*) but with an amplified magnetic field.

it will break, and move the ends to the poles, so that the original loop is deformed into a uniform tube which is parallel to the initial tube (fig. 2 e). At the loop formation the volume of the initial tube has decreased by $2\pi^2 R^3$ so that as the length is unchanged its cross-section has diminished by factor $\frac{2\pi R}{L}$. As the axial flux remains the same, the field has increased by a factor $1 + \frac{2\pi R}{L}$. If the tube deriving from the loop is added to the initial tube, the result is a tube with the initial volume and shape but with a field which has increased by $1 + \frac{2\pi R}{L}$. An iteration of the process will bring us up to any value of H_a.

The dynamics of a field amplification

5. In the preceding paragraphs it has been shown that an amplification of a magnetic field within a conducting liquid is kinematically possible. (This means also that the breeding of new lines of force is no really serious problem.) What remains in order to explain the generation of the terrestrial and solar magnetic fields is to find the dynamics which produces the postulated motions.

The dynamical problem is much more difficult because the behaviour of a set of electric currents in a fluid leads to mathematically very complicated problems, which have not been solved so far. By analogy with current-carrying wires certain conclusions could be drawn but they are only tentative.

Two parallel wires attract each other if their currents flow into the same direction. If the wires make an angle they tend to move so as to become parallel. A single loop of arbitrary form tends to expand and if free it will form itself into a circle. If a circular current is given, magnetic dipoles in its neighbourhood tend to move to the surface of the circle and place themselves in such a position as to amplify the magnetic field. If a multitude of current-carrying wires are placed near together they tend to move into such positions as to make the resulting dipole moment a maximum.

If the same laws hold for currents in a conducting liquid—but this could be established only after rigid calculations—currents in different parts of the liquid should adjust themselves so as to amplify each other. Eventually the result may be that the dipole moment becomes a maximum. Hence if only new lines of force are bred, they would move into such positions as to amplify the already existing field.

6. In §§ 3 and 4 we have only treated the magnetic lines of force without mentioning the currents producing them. In the ideal case of a single tube of flux the field is produced by currents flowing around the tube in an infinitely thin surface. In the case of a twisted flux tube there is also a current parallel to the tube. In the case of a uniformly twisted tube of force discussed in § 4 a current with homogeneous density $i = H_T/r$ flows inside the tube, the circuit being closed by a return current in an infinitely thin layer at the surface of the tube.

A set of force tubes of this ideal type will not interact. In reality, however, the currents spread in the environment, producing a magnetic coupling between the different parts of the liquid. This gives rise to mechanical forces which change the state of motion of the liquid. Accepting the analogy with the current-carrying wires we should expect the net result to be that the currents amplify each other.

Generation of magnetic fields in celestial bodies

7. We shall now try to apply what has been stated above to the generation of the general magnetic fields. We consider the liquid core of the earth as the place where the earth's magnetic field is generated. This core has a radius of about half the earth's radius. In the case of the sun the whole sun is gaseous so that the same considerations could be applied to the whole sun. In a liquid with the conductivity σ we have

$$\vec{i} = \sigma \left(\vec{E} + \frac{1}{c} \vec{v} \times \vec{H}\right) \tag{1}$$

$$\text{rot } \vec{E} = -\frac{1}{c} \frac{\partial \vec{H}}{\partial t} \tag{2}$$

Combining these equations and integrating along a closed line we find

$$\frac{\partial \Phi}{\partial t} = \int (\vec{v} \times \vec{H})\, \vec{ds} - c \int \frac{\vec{i}}{\sigma} \vec{ds} \tag{3}$$

where Φ means the magnetic flux through the surface limited by the line, and ds the line element.

The first term at the right hand side of (3) means the net flux transported by hydro-

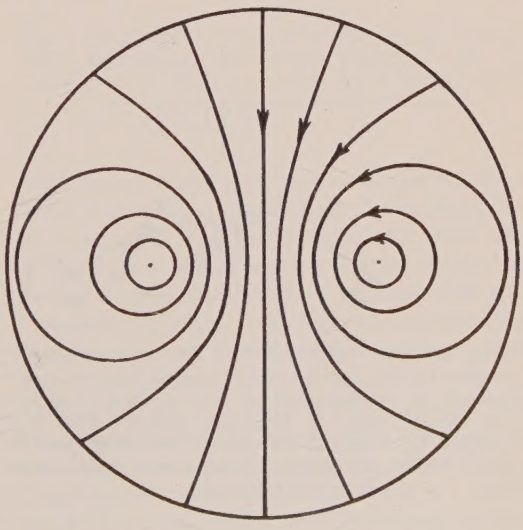


Fig. 3. Simple model of the magnetic field in the interior of the earth (or sun) with meridional field only.

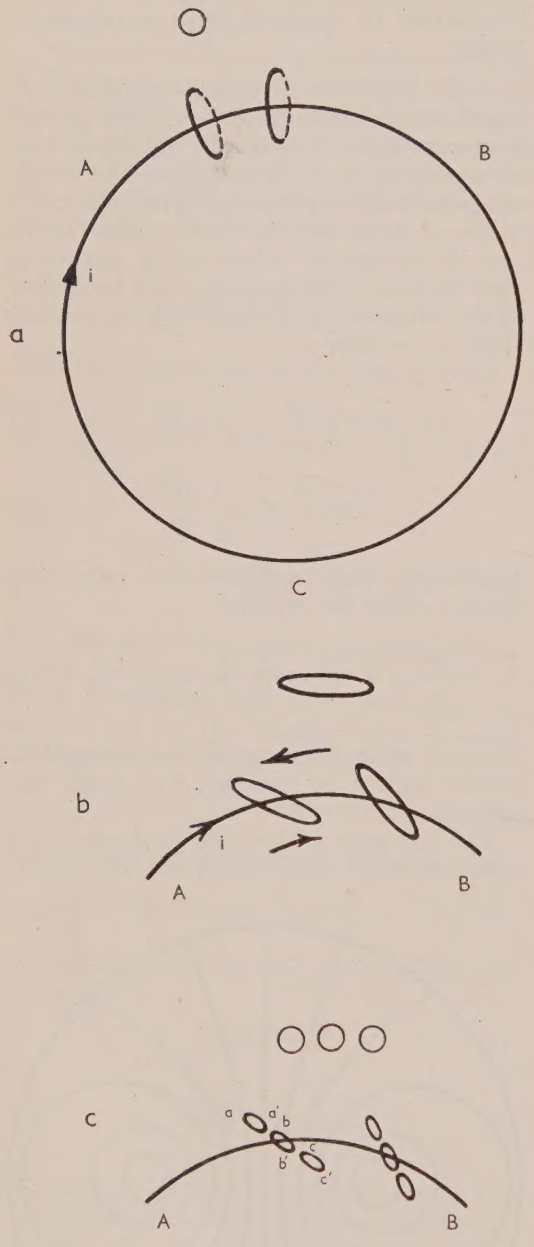


Fig. 4. Amplification of terrestrial field by means of the process of fig. 1. The neutral line ABC (fig. 4 a) is surrounded by circular magnetic lines of force (cross-section shown above). A differential hydrodynamic flow indicated by arrows in (*b*) deforms the circles into ellipses. These split up for example into three almost circular lines (*c*).

dynamic motion over the line of integration. If the flux Φ shall not diminish this transport must at least compensate the second term to the right, which denotes the consumption of flux due to the finite conductivity.

If we integrate over a circle in the equatorial plane of the earth, the net flux within this circle is directed southward. In order to amplify the field we must have a transport of southward flux into the circle, or a transport of northward flux out of the circle, or both.

Fig. 3 shows the simplest possible model of the field in the terrestrial core or in the sun. The field has meridional symmetry and is produced by a current which is more or less concentrated in a circle in the equatorial plane. Near this circle there is a circle called the "neutral line" where the magnetic field is zero.

If our integration line coincides with the neutral line there could obviously be no transport of flux over the circle because $H = 0$. Hence in this model the magnetic field cannot be maintained or amplified in a stationary state.

This difficulty, which has been stressed by COWLING [5], could be overcome by making the model asymmetrical and non-stationary. In fact, a neutral line $H = 0$ exists only because of the assumed symmetry. In a magnetic field of arbitrary form there exist singular points $H = 0$ and the neutral curve is not a continuous line but an assembly of such singular points.

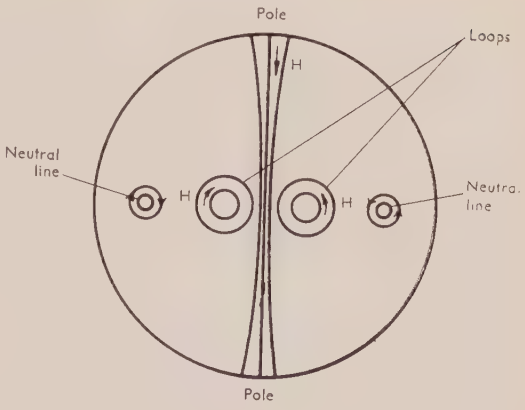
The possibility of a continuous transport of flux over a given circle in the equatorial plane is connected with the breeding of new lines of force. Without such a breeding the transport could only go on until all flux going southward through the equatorial plane has been brought within the circle and all northward flux has been brought outside it. Thus in order to maintain or amplify the field it is necessary to breed northward flux inside, or southward flux outside the circle, or both.

8. In order to study the process more in details we start from the model of fig. 3. Admittedly it has been shown that this cannot be exactly correct but it is useful as a first approximation and the necessary deviations may be considered as disturbances.

The breeding of new lines of force may go on in the whole volume of the terrestrial core and of the sun but the processes are somewhat different in different parts. In this paragraph we shall consider a process which may be active near the neutral line whereas in the next para-

graph we shall consider the breeding near the axis.

In the absence of a tangential field the lines of force near the neutral line are circles around this line (fig. 4 a). If the angular velocity of a tangential motion in the equatorial plane differs outside and inside the neutral line, the lines of force will be dragged out (Fig. 4 b). In this way mechanical motion is converted into magnetic field energy. If the line of force is prolonged very much it may divide into two or more loops. Speaking of the currents the dragging out corresponds to the production of a radial sheet of current due to the polarisation field $\vec{E} = \frac{1}{c} \vec{v} \times \vec{H}$ produced by the differential motion. The splitting of the prolonged loop means that the current density in the sheet is no longer uniform. It can be shown that an infinite sheet of current in a conducting liquid is unstable and will have a tendency to break up into separate lines of current. In this way southward lines of force are created at a' outside the line of integration ABC and northward lines of force are created at c inside the same line. A transport of the region a' inwards and of the region c outwards will give an increase of southward flux inside the line of integration. How this transport is effected dynamically is very difficult to analyse. The breaking up of the extended lines of force into smaller loops corresponds to the main current along the line ABC being twisted. If this twisted current is reshaped into a circle the expected transport is effected. It seems reasonable to assume that the current has a tendency to return to the circular shape. It should be observed that the breaking up of an extended line of force occurs when the tangential field surpasses the order of magnitude of the meridional field.



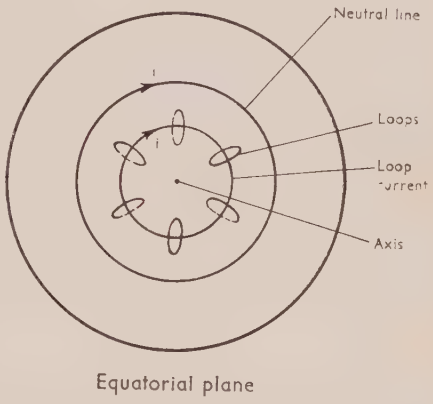


Fig. 5. Amplification of terrestrial field by the process of fig. 2. By twisting a tube of force loops are formed. The "loop current" of a symmetrical set of loops amplifies the initial current near the neutral line.

9. The breeding may also occur far away from the neutral line by means of the process outlined in § 4. This seems to be especially important near the axis. If the lines of force in this region are twisted by hydrodynamic motions, loops of magnetic flux will be formed. If a series of loops are bred symmetrically around the axis they will join to a toroidal field associated with a circular current in the equatorial plane. This current will expand due to its own magnetic pressure and also because of the attraction from the initial current near the neutral line. It is possible that the mutual attraction may prevent the initial current from expanding which it naturally should do because of its own magnetic pressure. If so the two current rings may coalesce so that there is a certain current (near the neutral line) which from time to time is fed by new rings. It is also possible that the general field derives from a series of expanding current rings, moving outwards in a long succession. In both cases there is a northward flux just outside the expanding currents which when passing a given line of integration effects the necessary transport of flux.

The Earths magnetic field

10. The process which we have considered is of a very general type. It should occur as soon as the average energy due to hydrodynamic motion within a part of the body exceeds the magnetic energy. The equality of these two sorts of energy is characteristic for the magneto-hydrodynamic waves. If such a wave is started by a hydrodynamic motion, half of the kinetic energy is transformed into magnetic energy.

Starting from a meridional field of a simple structure, the hydrodynamic motions may drag out the lines of force and effect a breeding of new lines of force directly, but a more obvious process is that the motions at first give the field a tangential component, which corresponds to a twisting of the lines of force. When this exceeds the meridional component new lines of force are bred and in this way the meridional field is amplified.

If the average field within a body is H the average motion which is necessary has the kinetic energy

$$\frac{\varrho v^2}{2} = \frac{H^2}{8\pi}$$

or

$$v = \frac{H}{\sqrt{4\pi\varrho}}$$

where ϱ is the density and v the velocity. For the earth's core we may put $H = 20$ gauss, $\varrho = 10$ which gives $v = 2$ cm/sec. For the sun $H = 100$, $\varrho = 1$ gives $v = 30$ cm/sec.

In order to maintain the general field at a certain strength the hydrodynamic motion must be supplied continuously with energy. ELSASSER and BULLARD suppose that the sourse of energy is nuclear reactions. Radioactivity in the earth's core produces a heating which gives rise to convection. In the solar core there is a very large release of nuclear energy.

Against this it may be objected that if these processes were the cause of the general fields we should expect a much larger difference in strength. The field at the surface of the earth's core can be estimated to about 5 gauss and this is of the same order as the field at the solar surface (10—20 gauss). Hence even the dipole moment per unit volume is of the same order. It is very unlikely that two fields of about the same strength should be caused by the release of nuclear energy, which is enormously much larger in the sun than in the earth. Admittedly it may be unfair to compare the earth's core with the whole sun because the magnetic field may be generated mainly in the solar core, but even so the difference in magnetic field would be very much smaller than the difference in energy production.

11. As an alternative there is the possibility that the energy is supplied by the non-uniform rotation which means that the ultimate cause is the braking of the rotating bodies by interaction with the environment. In the case of the earth the main braking is due to the tides, the action of which lengthen the day by 10^{-5} sec. per year. The tides brake the crust and if the braking is transmitted from the crust to the core by electromagnetic forces the energy released by the braking of the core may be converted into magneto-hydrodynamic energy. BULLARD [6] means that it is impossible that a differential rotation in the earth's core could be large enough to account for the magnetic field, because the magnetic field itself would force the core to rotate with about the same angular velocity. This conclusion depends upon what assumptions we make about the shape of the magnetic field in the core. If the field of the earth's core has the shape of fig. 3, that part of it from which magnetic lines of force go out to the crust, must rotate with almost the same velocity as the crust, according to FERRARO's theorem [7]. That toroidal part of the core, however, from which no magnetic lines of force go out to the crust has no direct electro-magnetic contact with this and may very well have a different angular velocity.

Suppose that the terrestrial dipole moment is a and the core has the radius R. If within the core there is a fairly uniform current density i in the tangential direction, this is of the order

$$i = a\, R^{-4}$$

The Joule energy developed per second by this current is of the order

$$p = i^2\, R^3\, \sigma^{-1} = a^2\, R^{-5}\, \sigma^{-1}$$

where σ is the electric conductivity. This should be compared with the energy developed in transferring rotational energy from the core to the crust. In order to find the order of magnitude, let us suppose that the core rotates as a rigid sphere thus neglecting that a part of

it is magnetically bound to rotate with the same velocity as the crust. The moment of inertia be I_1 and the angular velocity ω_1, whereas the corresponding quantities of the crust are I_2 and ω_2. When the angular momentum $I_1 d\omega_1 = -I_2 d\omega_2$ is transferred, the energy

$$dW = -\left[d\left(\frac{I_1 \omega_1^2}{2}\right) + d\left(\frac{I_2 \omega_2^2}{2}\right)\right] = \\ = (\omega_2 - \omega_1)\, I_1\, d\omega_1$$

goes into heat. If the "friction" is due to electro-magnetic forces, this energy is at first converted into magnetic energy, which finally is dissipated into Joule heat. Supposing that most of the magnetic energy is used to produce the general magnetic field we may put

$$p = \frac{dW}{dt}$$

or

$$a^2\, R^{-5}\, \sigma^{-1} = I_1\, \omega' \frac{d\omega_1}{dt}$$

where

$$\omega' = \omega_2 - \omega_1$$

is the differential angular velocity between the crust and the core. Putting $I_1 = \frac{2}{5}\, \varrho\, R_1^5$ (ϱ = density) we find

$$\omega' = \frac{5}{2}\, \frac{a^2}{\sigma\, \varrho\, R_1^{10} \frac{d\omega_1}{dt}}$$

We put $a = 8 \cdot 10^{25}$ gauss cm^3, $\varrho = 10$ g cm^{-3}, $R_1 = 3.5 \cdot 10^8$ cm. For $\frac{d\omega_1}{dt}$ we choose $3 \cdot 10^{-22}$ sec^{-2} which is the value of the retardation of the crust due to the tides, thus assuming that a stationary state is reached so that the core is retarded at the same rate as the crust. Accordind to BULLARD we have $\sigma = 3 \cdot 10^{-6}$. This gives

$$\omega' = 7 \cdot 10^{-8} \text{ sec}^{-1}$$

which corresponds to an excess rotational velocity of the core of about 0.1 %. There seems to be no certain information about the non-uniform rotation of the earth's core so this value cannot be checked at the present stage but does not seem unreasonable.

Thus even the braking of the earth suffices to maintain the terrestrial field. It is difficult to judge whether the energy is supplied in this way or by nuclear reactions or in both ways.

The difference between slow and rapid motion

12. If a force acts perpendicular to a string during a time which is short compared to the period of oscillation of the string a wave is produced which travels along the string. On the other hand, if the force is applied smoothly and acts during a time much longer than the period, no wave is formed but the string is statically deformed. The same holds for a magnetic line of force. A rapidly changing disturbance produces a magneto-hydrodynamic wave but a static or semistatic force produces a constant or slowly varying change in the shape of the line of force. It is likely that the factors producing the general magnetic field vary very slowly compared to the time of travel of magneto-hydrodynamic waves through the earth's core or through the sun. Considering the magnetic lines of force as strings, in this case we have not actually to do with the oscillations of these strings but with the forces stretching the strings.

In this way the production of the general magnetic field and the production of sunspot waves in the sun may go on rather independently, the former being due to slowly varying forces and the latter to forces of pulse character. This problem can be settled only when there is a general theory of the internal circulation taking account also of electro-magnetic effects.

Finally the relation with the magneto-hydrodynamic theory of sunspots should be discussed. In the development of this theory it was supposed that the magnetic field in the interior of the sun was homogeneous inside a sphere R_1 ($\sim 1.8 \cdot 10^{10}$ cm) and a dipole field outside R_1. This model is not very remote from the model we have discussed here. It should be observed that a superimposed tangential field does not affect the theory of sunspots because the velocity $\vec{V}$ of the magneto-hydrodynamic waves is proportional to $\vec{H}$ so that the motion in the meridional plane de-

pends only on the meridional field component.

A loop formation in the solar core must affect the motion of sunspot waves. Whether this has any connection with the sunspot cycle cannot be decided at present.

I wish to thank Mr. Herlofson and Mr. Lundquist for many discussions.

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Late Cenozoic Climatic Changes as Recorded by the Equatorial Current System

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Abstract

A method has been developed for using deep sea deposits as records of oceanic productivity and its changes. The sequence of sediments below the Equatorial Current system indicates a marked periodical change in productivity with time. This periodicity is assumed to correspond closely to the world-wide climatic changes during late Pliocene and Pleistocene. It is thought that the method offers possibilities for an accurate and detailed study of the history of the late Cenozoic climate. An explanation is given of the reaction of the Equatorial Counter Current to climatic changes.

Since it became clear during the latter part of the last century that the present geological period is characterized by alternating cold and warm periods, intensive work has been carried out in order to reveal the details of climatic evolution. Several difficulties have been met with, among them is the fact that the climatic variations themselves have often disturbed the type sequences and thus distorted the records. This is the case especially in a glacial environment. In these regions of the continents where the climatic changes have not caused extremely marked changes of facies their effect has mostly been so complicated that it is difficult to interpret the variations in terms of climatic changes.

It is also difficult if not impossible to find on the continents, on their shelves or on oceanic islands, localities where all factors tending to change the environment except the climatic factor have been constant throughout Pliocene and Pleistocene time. Tectonic and volcanic disturbances are the most important ones in this respect.

The ideal sedimentary profile for investigation of the late Cenozoic climatic changes should be situated far from the polar regions within which the violent facies changes interfere with the quantitative relationships. The environment of sedimentation should, moreover, be sufficiently extensive that the local influences can be easily recognized or be negligible.

Sediment profiles in such environments were first sampled by the German Deep Sea Expedition with the »Meteor». The pioneer work in deep sea biostratigraphy was carried out on this material by W. SCHOTT (1935) who was later followed by other workers, especially in the USA.

The present author's investigations on sediment cores, raised from the bottom of the Eastern Pacific during the Swedish Deep Sea Expedition 1947—1948, show that almost ideal conditions can be found in some, but not all, bathyal environments of sedimentation. The best relation between facies change and climatic change has thus been found in the equatorial region. This is easily understandable as the smallest displacements of facies bound-

[1] Kagghamra, *Grödinge*.

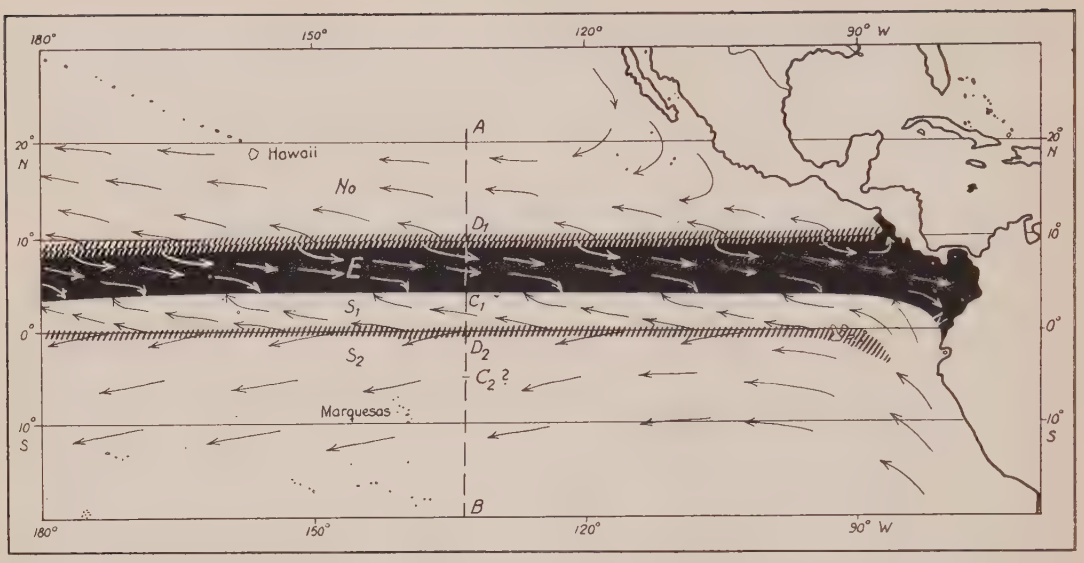

Fig. 1. The equatorial current system in the eastern Pacific. Arrows indicate schematically the direction of the surface currents. No North Equatorial current, E Equatorial Counter Current, S_1 northern and S_2 southern cell of the South Equatorial Current, N El Niño. Striped surfaces show distribution of cold upwelling deep water along the divergences. The figure refers to conditions prevailing during northern summer. D_1 and D_2 divergences, C_1 and C_2 convergences. C_2 is questionable and has hitherto not been recorded by oceanographical investigations. The possible existance of a secondary convergence at about 5° S is however indicated by the petrographical characters of the bottom deposit at this latitude.

aries will take place along the meteorological equator if simultaneous climatic changes are assumed in the Northern and Southern Hemispheres. The smallest effect of local disturbances has been found far from land in purely pelagic environments.

For the recording of the climatic events an additional factor is, however, necessary, viz. a sedimentation machine driven by an agent depending directly upon the global atmospheric circulation. The equatorial current system acts as such a machine and is maintained by the trade winds and thus, ultimately, by the thermic potential decrease between the equator and the poles.

Because of the important rôle of the Equatorial Current System, a brief summary will be given below of the dynamic relations, especially with regard to the Counter Current.

Strong currents are restricted to the surface layer of the oceans, whereas the movements of the deep water are believed to be very slow. In tropical regions the surface layer is heated by the radiation of the sun and the specific gravity of the water is therefore low. This lighter and warmer surface water is spread out over the deep water which is characterized by its high density due to the low temperature. The supply of nutrients such as phosphates, nitrates, silica, etc. is rapidly used up by the phytoplankton in the surface layer and transported down into the deep water as the dead organisms settle. The stable stratification prevents a rapid restitution of the nutrients to the surface through convection, and tropical waters are thus often characterized by a low plankton production and high transparency.

The trade winds drive before them the surface water thus giving rise to the Northern and Southern Equatorial Currents. Owing to the transport action of these currents, water is accumulated at the western borders of the oceans and a westward ascent in the ocean level is maintained in the tropical region. The accumulated masses of water find their way back to the East within the calm belt, thus forming the Equatorial Counter Current (fig. 1). As a result of the unequal distribution of the continents between the northern and southern hemispheres the line of symmetry

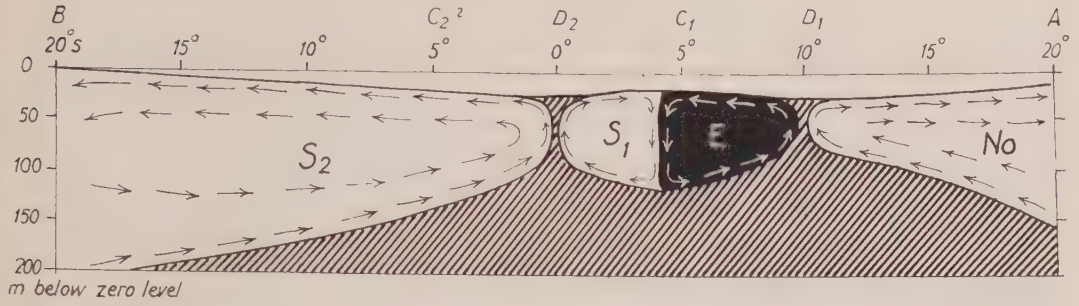

Fig. 2. Section through the equatorial current system along the meridian A—B in fig. 1. Indications as in fig. 1. Striped surface indicates the deep water body with tongues of upwelling. Zero level is arbitrarily chosen. Partly after SVERDRUP, JOHNSON and FLEMING (1946).

of the trade system and of the calm belt is displaced several degrees north of the geographical equator. Because of this the friction along the southern border of the current will be stronger than that along the northern border. An individual mass of water at the surface of the Counter Current will, therefore, not move straight eastwards but move over to the southern border of the current where it dives downwards and moves to the northern border on the under side of the current (fig. 2). A gyratory motion is thus superimposed on the eastward motion of the Counter Current, and owing to friction the adjacent masses of water of the Equatorial Currents are put into a similar but oppositely directed gyratory motion.

The existance of this mechanism is of decisive importance in the production of plankton in the equatorial region of the ocean. The upwelling along the divergences brings cold deep water rich in nutrients up to the surface, where the radiation from the sun makes an extensive assimilation possible. A strong production of plankton takes place, therefore, in the narrow divergence belts which run in the direction east—west across the oceans associated with the Equatorial Current System.

Further details of the dynamics and production of the present Equatorial Counter Current may be found in surveys and bibliographies such as in SVERDRUP, JOHNSON and FLEMING (1946). Detailed investigations based upon the material collected during the Swedish Deep Sea Expedition are at the moment being carried out by the hydrographers of the expedition, Dr. N. G. Jerlov and Dr F. Koczy.

The conditions for production of organic matter in the surface layer of the ocean greatly influences the bottom deposits. MURRAY and RENARD (1891) and REVELLE (1944) have pointed out the agreement between the path of the Equatorial Counter Current and the northern border of the Pacific Globigerina ooze area.

The present investigation of sediment cores from the Eastern Pacific confirms the correlation between distribution of currents and type of deposit. The biogenous component and simultaneously the rate of sedimentation increase strongly below the convergence C_1 (fig. 1 and 2), where a biolith, rich in fossils of foraminifera, radiolaria and diatoms is thus deposited.

The fact that the sedimentation is most intensive below the convergence in spite of the greatest production being found in the divergences (GRAHAM 1941) may be explained by fig. 2. The phytoplankton produced in the divergences D_1 and D_2 is assumed to be carried to the convergence C_1 where it is withdrawn from the euphotic zone, dies and settles (cf. JOHNSON in SVERDRUP e. a. 1946). The biogenous sedimentation also increases at latitude 5° S which might possibly indicate a secondary, hitherto unknown convergence at this latitude.

The abundant supply of organic matter below the Counter Current makes possible the existence of a comparatively rich bottom fauna. The activity of the mud eating animals is easily recognized in the sediment in the

form of crushed shells (ARRHENIUS 1950 a), skeletal remains, coprolithes, digging structures, etc.

North of the convergence (C_1 in figs. 1 and 2) of the Equatorial Counter Current, the share of the biogenous component in the sedimentation decreases and the fossil-rich biolith turns into a clay, poor in fossils and with a low intensity of sedimentation (0—0.4 cm per 1,000 years). The boundary between the calcium carbonate ooze and the red clay area is astonishingly sharp and the diffusion caused by lateral transport of settling particles and by the seasonal changes in the path of the current system is obviously not big enough to disturb the general distribution pattern.

The sediment cores offer an opportunity for studying the changes in the Equatorial Current System during the Pliocene and Pleistocene epochs. For the recording of the displacement of the current path and of changes in the intensity of production the author has chosen to measure the absolute content of dominating fossils, integrity degree of foraminafera and size distribution of certain radiolaria (compare WEIBULL 1949) in the sediments, and the content of carbonates, humus carbon, nitrogen and phosphorus. Approximately 2,000 samples have been extracted, representing on an average each ninth centimeter of the East Pacific cores. Of chief interest here are the results of the quantitative microfossil investigations, as these clearly indicate a rhythmical change in plankton production with time. The concentration in the sediment of foraminifera, for example, shows that the present production is fairly low. In subrecent layers the concentration is higher and if we move backwards in time we will pass a marked maximum, a layer with decreasing concentration and then a minimum, obviously caused by a still lower production than the present one. Then follows a series of five maxima and minima of considerable amplitude and thereafter two maxima of a smaller order of magnitude. The two youngest maxima are the greatest, and both are complex with four to five subordinate peaks.

Of the minima the third from the top (the present one included) is the most marked. The minima are also often complex, the next youngest one shows, for example, a small but marked subordinate maximum. This stratification can be traced in an east — west direction over the whole area investigated, comprising approximately 4,800 km and the observations made by the author during the expedition indicate that the phenomenon is a global one. However, within the equatorial biolith belt marked latitudinal changes are obvious. In isochronous layers the highest concentrations are met with below the convergence of the Equatorial Counter Current. A tentative explanation of this has been given above. The greatest changes in concentration of shells with time are also found below the zones of downwards movement of the surface water. The content of shells of foraminifera $>150\,\mu$ varies in this environment between approximately 900 and 30,000 shells per gram of air dry sediment.

Below the ten meter thick layer series which is characterized by strong variations in the supply of plankton, the content of shells decreases to 500 to 1,000 shells/g sediment and the marked stratification mentioned above disappears. These lower layers are believed to be of late Pliocene age.

North of the convergence of the Counter Current the concentration of fossils in the sediment decreases rapidly as has already been mentioned. A sediment core at latitude N 8°25′ longitude W 128° 48′ consists of a red clay, almost free from calcareous fossils. At a depth of 380 cm the clay changes gradually to a calcareous sediment with several hundred shells per gram. According to Dr. F. Brotzen, the benthonic foraminifera indicate that this fauna is very likely to be Tertiary. Stratigraphically these layers are associated with the series mentioned above, which as assumed to be late Pliocene. At latitude N 13° 17′, longitude W 127°25′ a ten meter long core appeared to consist entirely of red clay, poor in fossils. The Tertiary area with a comparatively high plankton production has thus a northern border somewhere between N 8° 25′ and N 13° 16′. This limit is approximately parallel with the border of the present production belt.

In this connection it may be mentioned that a secondary dissolution of calcareous shells in red clay facies should probably not be ascribed the great importance often given it. Secondary dissolution influences all the more the final distribution of radiolaria and diatoms, especially in sediments low in calcium car-

bonate. The final distribution of planktonic foraminifera in bathyal deposits is believed primarily to depend upon the production in the surface layer of the ocean and upon the crushing by mud eating bottom organisms.

The general sequence of the Pacific deep sea deposits presented above indicates that the equatorial production belt of the ocean surface was comparatively broad during late Tertiary time but produced farily small amounts of plankton. The transition to Pleistocene time involved the Counter Current becoming compressed and that the intensity of production increased. This effect is believed to have been caused by an intensification of the trade winds. Such an intensification ought to have caused an increased transportation of water by the Equatorial Currents and by the Counter Current and therefore an increased upwelling in the divergence belts. If this was in fact the case the strong rhytmic changes in plankton production recorded in the Pleistocene part of the sequence ought to have been caused by changes in the strength of the trade winds and thus ultimately by periodical variations in the global distribution of atmospheric temperature and pressure.

The suggestion has been made elsewhere as an explanation of the obvious stratification also observed in equatorial areas other than the Eastern Pacific — that the Equatorial Counter Current was subjected to considerable latitudinal displacements during the Pleistocene climate changes. This hypothesis is clearly contradicted by the results from the investigation of the East Pacific sequences, which indicate great constancy in the path of the Counter Current throughout Pleistocene time. The stability of the location of the current is explained by the theory outlined above regarding the reaction of the Counter Current to simultaneous intensity changes of the northeast and southeast trade winds. Reinforced trades will ultimately have a compressing effect upon the Counter Current, but this effect will be increasingly counteracted by the swelling of the current due to the more intensive transportation of water. At a certain intensity of the trade winds equilibrium is practically attained, where an increase in intensity does not markedly influence the breadth of the current. This approximate equilibrium appears to have been established during the climatic change at the transition between the Pliocene and Pleistocene epochs. It is very likely that smaller displacements of the current system took place later, but they are too insignificant to be recorded by the net of stations of the Swedish Deep Sea Expedition. The lack of significant displacements of the Counter Current during Pleistocene time appears to be a definite proof that the climatic changes were synchronous in the Northern and Southern Hemispheres. The results presented here are thus in conflict with the heat distribution theory of climatic changes put forth by Milankovitch e. a. (see for example survey in ZEUNER 1945). If the Simpson hypothesis (SIMPSON 1934, 1940) is applied to the present results, each major production period should correspond to two glacial and two interglacial ages. Simpson's theory would thus demand twelve or more glacial periods. At present it seems more probable, therefore, that each glacial period corresponds to one single period of intensified atmospheric circulation.

What has been said above regarding the stability of the Counter Current's location during quarternary refers only toopen ocean areas where the current runs undisturbed by continental coasts and archipelagos. Off the South American coast the periodical intensifications of the Counter Current's transport of water have probably resulted in corresponding expansions of El Niño (fig. 1) to the south. The material of the Swedish Deep Sea Expedition does not, however, include cores permitting a study of such changes.

Changes in the species distribution of planktonic foraminifera recorded in pelagic equatorial deposits have hitherto been interpreted as caused by cooling of the surface water due to decreased radiation from the sun or to invasion of cold water from higher latitudes. The present results make it probable that changes in the temperature of equatorial surface water in the oceans may be largely due to changes in the upwelling of cold deep water. Changes in distribution of planktonic species will, therefore, be a complex result of variations in temperature and in the supply of nutrients.

Strong trade winds ought to indicate a great decrease in temperature from the equator to the poles and as a result hereof intense atmos-

pheric circulation, whereas faint trade winds indicate the contrary conditions. Thus it appears to be highly probable that the changes in production in the Equatorial Current System correspond closely with the worldwide climatic changes, which characterize especially the Pleistocene epoch. Superior and subordinate maxima should in this case correspond to ice ages and subages whereas minima should indicate interglacial ages and subages. If this interpretation is correct, petrographical and paleontological investigations of pelagic bathyal deposits influenced by the Equatorial Current System should permit a much more detailed study of the history of the Late Cenozoic climate than has hitherto been possible.

This paper is to be regarded as preliminary. The complete presentation of the East Pacific material including analyses and stratigraphical correlations will follow in the Reports of the Swedish Deep Sea Expedition.

The programme for working on the East Pacific investigation has been summarized elsewhere (Arrhenius 1950 b).

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Photographic Emulsion Method Applied to the Investigation of Radioactive Deposits from Atmospheric Air

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(Manuscript received 6 December 1949)

Abstract

This article describes a combined activation-photographic detection method applied in the investigation of radioactive deposits from atmospheric air. It is shown that with this simple and convenient method it is possible to obtain many fine details of the phenomenon also in case of quite weak activity. It is also shown that the method can be employed with success in the investigation of the mobility of radioactive carriers in atmospheric air.

Introduction

The incentive to this investigation arose out of observations carried out in connection with the introduction of measurements of small-ion density of atmospheric air at the Institute of High Tension Research, University of Uppsala (NORINDER a. SIKSNA 1949). An abnormally high ion-density was observed in the rooms of the Institute building (50,000—60,000 n/cm^3 in the transformer-room, 3,000—20,000 n/cm^3 in other rooms; in the open air in the environments of the Institute the ion-density was usually 900—1,500 n/cm^3, more rarely it increased to 3,000—3,500 n/cm^3). It was assumed that the source of the higher ion-density in the Institute building was the transformer-room, and the cause of it—the radioactivity of the air.

Therefore, a method was sought with which it would be possible to carry out investigations of the radioactivity of atmospheric air without incurring too great expenses.

The photographic emulsion method seemed to us to meet these requirements. In recent times this method has found large application in many investigations (POWELL and OCCHIALINI 1947, YAGODA 1949), but as far as we know, it has not yet been applied to the investigation of the radioactivity of atmospheric air.

Method employed

The technique used was as follows: a freely suspended body is connected to the negative terminal of a high tension rectifier and exposed in the room where the measurements are to be made. After several hours of electrical exposure photo-plates are placed on the plane surfaces of the body with their emulsions touching the metal surfaces, after which photo-exposure follows. After the development of the plates the tracks of α-particles produced in the emulsion are measured microscopically.

The method employed can be regarded as the activation method of ELSTER and GEITEL (1902, 1903) combined with the photo-emulsion method for the determination of induced activity.

With this simple and convenient combined *activation-photographic detection* method it is possible:

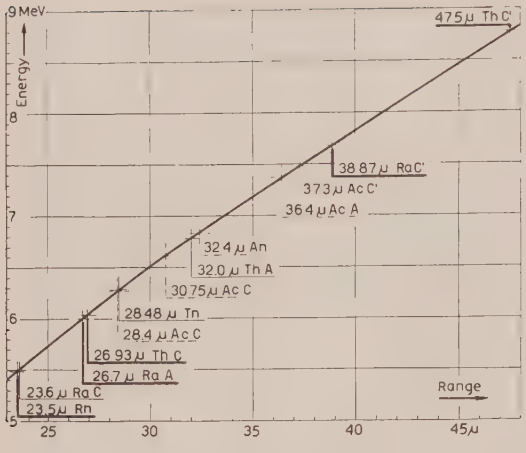

Fig. 1. Range-energy curve for α-particles of radon (Rn), thoron (Tn) and actinon (An) and their decay products in Ilford "Nuclear Research" emulsions.

1. to determine individual tracks of the radioactive deposit from atmospheric air,
2. to identify from the ranges the origin of individual α-particles; the range-energy relation of α-particles of emanations and all their decay products is given in fig. 1 (Lattes, Fowler, Cuer 1947),
3. to obtain many of the details of the phenomenon also in cases of quite weak activity,
4. to derive conclusions concerning the relation of contents of different kinds of active atoms in the atmospheric air from the relation of number of tracks.

Experiments

Corresponding to the method outlined above experiments were carried out according to the following arrangement:

1. The test body—a brass disc, 1 cm thick and 5 cm in diameter, with rounded edges—was connected with the negative terminal of a high tension rectifier (4.5 kV) and exposed in the room where it was to be tested.

Activation in such a manner was carried out in the transformer-room, in a work-room on the first floor of the Institute building and in the open air in one of the observation huts which had been built especially for the ion-density measurements at the Institute. The duration of the electrical exposure was 8—11 hours. After the electrical activation photo-plates were placed on plane surfaces of the brass disc with the emulsion layers in contact with the metal surfaces, whereupon photo-exposure followed. The photo-plates employed were Ilford "Nuclear Research Plates" Type E 1 with a 50 μ emulsion. The duration of the photo-exposure was 60—135 hours.

2. On the inside electrode of an Ebert ion-aspirator, connected permanently with the negative terminal of a dry battery (117 V respectively 240 V), a thin paper hull was slid. By aspirating with an air-stream through the aspirator, the hull was then activated for 9—11 hours. After this electrical exposure the hull was removed cautiously from the electrode, pressed flat between the emulsions of two photo-plates and held in that position for photographic exposure.

3. Already on the first plates exposed tracks could be traced which were identified as originating from Rn α-particles. Therefore the brass disc was grounded and exposed in this state in the transformer-room with a view to examine the possibility of obtaining an activation by the occlusion of radon in the metal without an electrical field.

Results

With the method described above the following has been established:

1. The absence of anticipated RaA—RaC' stars. The explanation of this fact may, perhaps, be understood from the following consideration: when a RaA-atom, situated on the surface of an activated body, has ejected its α-particle towards the emulsion it will be moved by the recoil into the body and the α-particle ejected afterwards from RaC' would not be directed towards the emulsion.

2. Stars of successive decay $Rn \xrightarrow{\alpha} RaA \xrightarrow{\alpha} (RaB) \rightarrow \rightarrow (RaC) \rightarrow RaC' \xrightarrow{\alpha}$. 1—2 % of all measured tracks are such radon stars, with the starting points of the stars located 10—20—30 μ deep in the emulsion.

3. Range spectra by means of which it is possible to analyse active substances present. Figures 2, 3 and 4 contain examples of spectra of the ranges of α-particles as obtained in the present investigation. The spectra from the inside rooms of the Institute building (fig. 2) show a distinct maximum with R = 39 μ, which corresponds to RaC', a second with R = 27 μ, correspond-

ing to RaA, and a third with $R = 24\,\mu$, corresponding to Rn. The spectra obtained by activation in the open air (fig. 3) also show maxima which might be attributed to the above mentioned ranges of Rn, RaA and RaC′ α-particles, but they are more blurred since two more intensive maxima with $R = 47\,\mu$ and $R = 27\,\mu$ (corresponding to ThC′ and ThC) have been superposed across them. Similar spectra are obtained by activating the paper hull over the negatively charged inside electrode of the Ebert ion-aspirator. The spectrum (fig. 4) obtained when the test-disc was grounded and exposed in the transformer-room has also the three maxima for the Rn-series, but for Rn the maximum is greater than it was when electrically activated.

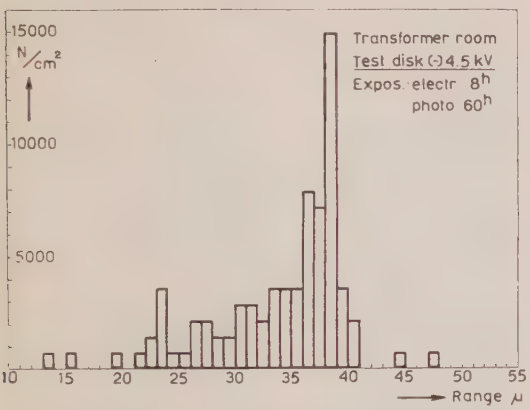

Fig. 2. Range spectrum of α-particles obtained from the transformer-room by activating the test-disc connected with a 4.5 kV negative tension.

4. Tracks of Tn decay products at very small rates of Tn in the atmosphere. It can be shown that also in those cases when the equilibrium state for the Tn series is not reached, the number of collected atoms of Tn decay products to be expected can be of the same order as that of the Rn products even with a ratio $Tn : Rn = 10^{-4}$ (Fig. 3), a number which may be accepted as a middle value in atmosphere. The reason for this is the difference in the half-lives of *Rn* and *Tn*, and of *RaB* and *ThB*.

5. Possibilities for detecting the presence of An decay products in the atmosphere. It can be shown that also for a ratio $An : Rn = 5.5 \times 10^{-7}$, estimated from the relative abundance of the two principal uranium isotopes U I and AcU

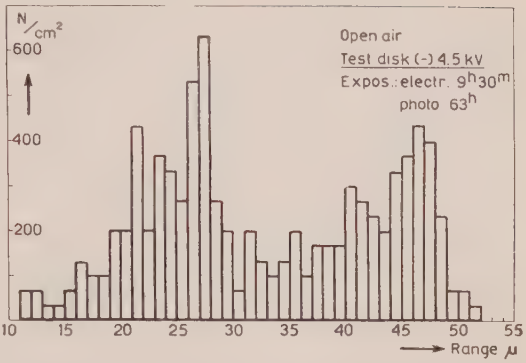

Fig. 3. Range spectrum of α-particles obtained in the open air by activating the test-disc connected with a 4.5 kV negative tension.

and their rates of disintegration, the ratio of the number of corresponding RaC and AcC atoms in equilibrium with their emanations will be RaC : AcC $\approx$ 200. In all the range spectra one can see a small maximum at 31—32 μ and, perhaps, another at 28—29 μ. It seems probable that tracks with these ranges may be interpreted as resulting from AcC, for in fig. 1 we have for AcC ranges 30.75 μ and 28.4 μ.

6. The presence of Rn atoms on the activated body. Direct arguments in favour of this are: the stars of successive decaying $Rn \xrightarrow{\alpha} RaA \xrightarrow{\alpha} \rightarrow (RaB) \rightarrow (RaC) \rightarrow RaC' \xrightarrow{\alpha}$ and the group of tracks with ranges around 23 μ; as indirect argument may also be considered the occurrence of a relatively large number of RaA tracks on several plates.

7. The deep penetration of active atoms into the activated bases. The spectral lines of range distribution are asymmetrical, spreading toward

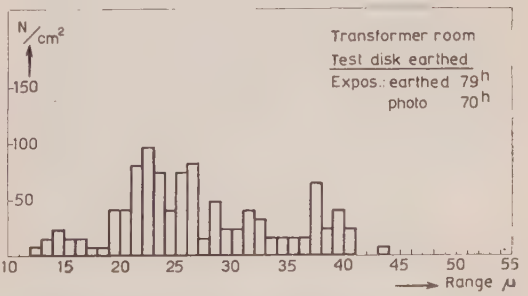

Fig. 4. Range spectrum of α-particles obtained from the transformer-room when the test-disc was grounded.

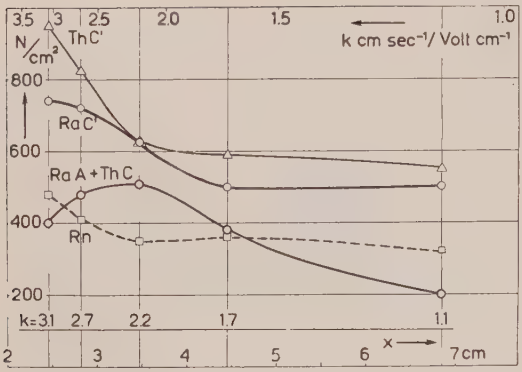

Fig. 5. Area density of a particular kind of active atoms deposited on the hull slid over the inside electrode of the Ebert ion-aspirator as a function of the distance x from the top end of the electrode. k — the limit mobility.

smaller ranges; one may suppose that many of the active atoms deposited must be hammered into the base by the electrical field or by the recoil, and that the α-particles ejected during the following decay therefore must follow a path through the material of the base.

8. The possibility to activate a hull slid over the inside electrode of the Ebert ion-aspirator applying the usual tension, in order to detect activity on the photographic emulsion. This convenient arrangement makes it possible to determine the mobility of the different active carriers in the atmospheric air and to distinguish between them. The area density of a particular kind of active atoms deposited on the hull slid over the inside electrode of the Ebert ion-aspirator, expressed as a function of the distance x from the top end of the electrode, is shown in fig. 5. We must desist from calculating the distribution of active carriers with respect to mobilities since this would require additional experiments in order to enlarge the measuring interval and to clear up several phenomena referred to above, for example the occlusion of radon. We have thereby wished to demonstrate that the method can be used for investigating mobility.

Acknowledgement

The author is very grateful to Professor H. NORINDER for his suggestion to investigate the radioactivity in atmospheric air, in connection with the observations made in our common investigation of small-ion density, for his liberal willingness to support this investigation, and for his interest throughout the study.

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The author likewise wishes to express his thanks to Professor A. LINDH for placing at his disposal Ilford "Nuclear Research" plates, and to Professor E. NORIN and fil. lic. M. FRIES for permitting the use of microscopes for the first measurements.

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Variation in the Earth's Angular Velocity Resulting from Fluctuations in Atmospheric and Oceanic Circulation[1]

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(Manuscript received 20 March, 1950)

Abstract

Fluctuations in the circulation of the atmosphere are associated with very small anomalies in the angular velocity of the earth. The seasonal component of these anomalies has been computed from weather maps, and is found to agree, with respect to magnitude and phase, with anomalies first reported by STOKYO in 1936 on the basis of astronomic observations. The effects of fluctuations in the oceanic circulation, and of shifting of air and water masses, have been estimated to account for not more than 15 per cent of the observed effect.

The detection of shorter period fluctuations is presently limited by the accuracy of astronomic time determination. The developments of an atomic clock and of the photographic zenith tube open the possibility of determining anomalies persisting for only two weeks, and of providing a new tool for obtaining an index of *integrated* atmospheric circulation.

Introduction

The problem is well stated by STARR (1948):

> "Although it is true that the combined system composed of the earth and the atmosphere cannot alter its total absolute angular momentum except for extremely slow secular changes resulting from tidal action as was pointed out by Darwin and others, still there is no reason to expect that the partition of angular momentum in the composite system should remain constant when seasonal and other short time-intervals are considered. Because of the great contrast in the moments of inertia of the two components, short-period anomalies of this kind represent major anomalies in the behavior of the wind system, but, on the other hand, imply practically undetectible inequalities in the rate of the earth's rotation."

WIDGER (1949) affirms that "there is little hope" of detecting these minute inequalities. Actually such inequalities have been reported, although not interpreted, by NICOLAS STOYKO, a French astronomer, in 1936, 1937, and 1949. Similar observations have been made by SCHEIBE and ADELSBERGER in Germany and by FINCH in England.

In a recent paper VAN DEN DUNGEN et al. (1949) have ascribed the fluctuations observed by STOYKO to seasonal shifts in air mass. Their computations (for the northern hemisphere only) account for 20 per cent of the observed effect, and the discrepancy is laid to uncer-

[1] Contribution from the Scripps Institution of Oceanography, New Series No. 472. This work represents results of research carried out for the Office of Naval Research under contract with the University of California.

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tainties in the various assumptions. We have recomputed this effect, taking into account the corresponding shift of air mass in the southern hemisphere (which largely compensates for the shift in the northern hemisphere), the local pressure variations rather than those reduced to sea level, and the tendency of the ocean water to redistribute itself under the influence of changing atmospheric pressure. Each of these factors substantially reduces the effect, and the computed anomalies in the earth's rotation resulting from the shift in air masses amounts to less than 2 per cent of the observed effect. The fluctuations in the angular momentum of the atmospheric circulation is, however, of the right magnitude to account for the observed effect.

Conservation of angular momentum

Neglecting secular changes resulting from tidal friction,

$$I_e\Omega_e + I_a(\Omega_e + \omega_a) + I_o(\Omega_e + \omega_o) = \text{constant}, \quad (1)$$

where I_e, I_a and I_o are the moments of inertia of the solid earth, atmosphere and ocean, respectively, Ω_e the earth's absolute angular velocity, ω_a and ω_o the components along the earth's axis of the *relative* angular velocities of atmosphere and ocean. Differentiation gives

$$-\frac{\Delta\Omega_e}{\Omega_e} = \frac{\Delta(I_a\omega_a)}{I\Omega_e} + \frac{\Delta(I_o\omega_o)}{I\Omega_e} + \frac{\Delta I}{I}, \quad (2)$$

where $I = I_e + I_a + I_o$

Angular momentum of the atmospheric circulation

The relative angular momentum of the atmosphere between the levels h_1 and h_2, the latitudes φ_1 and φ_2, and the longitudes Θ_1 and Θ_2, is given by

$$I_a\omega_a = R^3 \int_{h_1}^{h_2}\int_{\varphi_1}^{\varphi_2}\int_{\Theta_1}^{\Theta_2} \varrho u \cos^2\varphi \, dh \, d\varphi \, d\Theta, \quad (3)$$

where R is the radius of the earth, ϱ the density and u the zonal wind. The largest portion of the angular momentum is contained in the "jet stream" and surrounding westerlies, which extend approximately from 25° to 60° latitudes.

The values in Table 1 have been computed from weather maps by numerical integration along vertical sections from the surface to 17 km. For the northern hemisphere[3] sections along 100° E and 80° W longitudes were used. It is assumed that those sections are representative of conditions throughout 90° of longitude to either side. Only one section is available for the southern hemisphere (Loewe and Radok, 1950), but it must be realized that this section is far from typical for other longitudes. The easterlies have been taken into account in the northern hemisphere, but not in the southern hemisphere where the belt of easterlies is narrow and their effect comparatively small.

According to Table 1 the seasonal variation in the southern hemisphere amounts to less than one third the seasonal variation in the northern hemisphere. The reason for this is twofold: (*a*) the jet stream in the southern hemisphere during the southern winter is narrower than the northern jet stream in the northern winter; (*b*) the westerlies in the southern hemisphere during the southern summer are very much stronger than the westerlies in the northern hemisphere during the northern summer. The seasonal variation in the southern hemisphere may be even smaller than indicated in Table 1, since our calculations are based on geostrophic winds, and these are almost twice the observed zonal winds at Townsville, Brisbane and Sydney during the southern winter. No wind observations are available for the summer season.

Table 1. Relative angular momentum of the atmosphere in units of 10^{32} g cm^2 sec^{-1}. The northern hemisphere is divided in two parts, centering at stated longitudes.

	Northern Hemisphere			Southern Hemisphere
	100° E	80° W	Total	150° E
$I_a\omega_a$, January	6.8	5.9	12.7	6.9
$I_a\omega_a$, July	0.2	−0.1	0.1	10.7
$\Delta(I_a\omega_a)$	6.6	6.0	12.6	−3.8

Setting $\Delta(I_a\omega_a) = 9 \times 10^{32}$ g cm^2 sec^{-1}, $I \approx I_e = 0.8 \times 10^{45}$ g cm^2, $\Omega_e = 0.73 \times 10^{-4}$ sec^{-1}, gives a relative seasonal range of

[3] We are indebted to Dr Yale Mintz of the Meteorology Department at the University of California in Los Angeles for making available the mean monthly cross section on which our calculations for the northern hemisphere are based.

$$\frac{\Delta(I_a \omega_a)}{I\Omega_e} = 1.6 \times 10^{-8},$$

or about one part in sixty million.

Angular momentum of oceanic circulation

To estimate fluctuations in the angular momentum of the ocean circulation we have computed numerically the angular momentum for two of the largest ocean current systems (SVERDRUP et al., 1946, pp. 614, 710):

Antarctic Circumpolar Current:
1.0×10^{32} g cm^2 sec^{-1}

Equatorial Easterly Currents, all oceans:
0.5×10^{32} g cm^2 sec^{-1}.

The angular momentum of the entire zonal circulation of the oceans is perhaps 3 to 4×10^{32} g cm^2 sec^{-1}, and the range in fluctuations probably not more than 20 per cent of this value, or

$$\frac{\Delta(I_o \omega_o)}{I\Omega_e} \approx 0.15 \times 10^{-8}.$$

It may seem surprising that the ocean circulation should contribute only one eighth of that of the atmosphere, especially since the mass of the atmosphere corresponds to only 10 meters of water compared to a mean ocean depth of 4,000 meters. However the mean zonal velocity of the ocean water is probably less than 2 cm/sec, whereas the mean zonal wind speed in the westerlies (weighted for density stratification) is 20 m/sec. Thus the ratio in masses of 400 : 1 is more than offset by a velocity ratio of 1 : 1,000. An additional factor is the larger *percentage fluctuations* of the atmospheric circulation about its mean value.

Moment of inertia

The third term on the right side of equation (2) includes changes in the moment of inertia of the atmosphere, the ocean and the earth. Changes in the ocean are due partly to the effect of wind stress (to be considered later), partly to the effect of atmospheric pressure on the sea surface in redistributing the waters of the ocean. We shall consider two extreme cases: the case of immovable water, as if the entire ocean were frozen, and the case for which complete equilibrium has been achieved. For seasonal variations equilibrium is probably nearly attained. Very short period fluctuations fall into the first category.

For the case of immovable ocean water we have estimated ΔI_a according to the equation

$$\Delta I_a = \frac{R^4}{g} \int_{-\pi}^{\pi} \int_{0}^{2\pi} \Delta p \cos^3 \varphi \, d\varphi \, d\Theta, \qquad (4)$$

where Δ again represents January values minus July values. The mass of air above a unit area of the earth's surface was set equal to p/g, and p determined from normal monthly weather maps (U. S. Weather Bureau, 1946, and BRUNT, 1939, pp. 8—9). In equation (4) p is the *local* pressure, and not the corresponding value reduced to sea level, as it is given on climatic charts.[4] Accordingly it was necessary to remove the sea level correction from the p values on the monthly weather charts (JEFFREYS, 1929, pp. 239—244). Values of Δp for pressures reduced to sea level, and for local pressure variations, are shown in Table 2. The quantities $[\Delta p]$ in Table 2 represent averages around the stated latitude circles. In general the local pressure variations are smaller than corresponding sea level pressure variations, and in high latitude the two variations are opposite in sign. Consequently the seasonal changes in the moments of inertia are substantially reduced when allowance is made for the elevation of the meteorological stations. In the northern hemisphere the reduction is by a factor of 2.7, in the southern hemisphere by only 6 per cent. For both hemispheres $\Delta I_a = -0.83 \times 10^{36}$ g cm^2 compared to a value of *plus* 1.8×10^{36} g cm^2 obtained by VAN DEN DUNGEN et al. (1949).

The value of ΔI_a is even further reduced if the ocean water is permitted to move to attain equilibrium. We adapt a method given by JEFFREYS (1929). Let Δm designate the change (January minus July) per unit surface area of the *combined* mass of air and water. Conservation of mass requires that

$$\int_E \int \Delta m \, ds = \int_o \int \Delta m_o \, ds + \int_L \int \Delta m_L \, ds = 0,$$

[4] The error resulting from the use of sea level pressure has already been pointed out by Angot. See Annuaire de la Société Météorologique de France, T. 35, 1887.

Table 2. Computation of ΔI_a (pressure in mb).

	Latitude	Sea level pressure		Local pressure		Local pressure	
		$[\Delta p]$	$\int_0^{2\pi} \Delta p \cos^3\varphi \, d\Theta$	$[\Delta p]$	$\int_0^{2\pi} \Delta p \cos^3\varphi \, d\Theta$	$\int_L \Delta p \cos^3\varphi \, d\Theta$	$\int_L \Delta p \cos\varphi \, d\Theta$
Northern Hemisphere	80—85	1.1	0.0	— 0.7	0.0	0.0	— 1.7
	75—80	1.3	0.1	— 1.1	— 0.1	— 0.2	— 4.2
	70—75	0.7	0.1	— 1.7	— 0.3	— 0.3	— 3.3
	65—70	2.6	0.9	0.2	0.1	0.6	3.8
	60—65	3.4	2.1	— 2.2	— 1.4	— 0.3	— 1.5
	55—60	3.5	3.5	0.0	0.0	2.2	7.2
	50—55	4.1	5.9	— 2.2	— 3.2	0.9	2.5
	45—50	4.4	8.6	— 1.9	— 3.7	2.7	6.8
	40—45	4.9	12.3	— 0.5	— 1.3	4.4	8.0
	35—40	5.5	17.2	— 0.1	— 0.3	3.1	5.0
	30—35	6.3	23.8	0.1	0.2	— 0.4	— 0.6
	25—30	6.9	30.0	0.5	2.2	— 1.7	— 2.1
	20—25	6.4	31.7	5.7	28.3	18.7	22.0
	15—20	4.7	25.5	4.0	21.9	9.0	9.9
	10—15	2.3	13.4	2.2	12.6	3.7	3.8
	5—10	1.4	8.4	1.2	7.3	— 0.4	— 0.4
	0—5	1.4	8.5	1.4	8.7	1.1	1.1
	Σ		192.0		71.0	43.1	56.3
	ΔI_a in 10^{36} g cm^2		2.81		1.04	0.63	
Southern Hemisphere	0—10	— 2.7	— 17.1	— 2.7	— 16.8	— 5.8	— 5.8
	10—20	— 4.4	— 25.0	— 4.1	— 23.1	— 7.4	— 7.9
	20—30	— 5.4	— 24.6	— 4.8	— 22.1	— 6.0	— 7.4
	30—40	— 1.6	— 5.5	— 1.6	— 5.4	— 0.4	— 0.6
	40—50	2.0	4.5	2.0	4.5	0.1	0.1
	50—60	1.3	1.5	1.3	1.5	0.0	0.0
	60—70	— 4.0	— 1.9	— 4.0	— 1.9	0.0	0.0
	70—80	— 2.1	— 0.2	— 2.1	— 0.2	— 0.1	— 2.7
	Σ		— 68.3		— 63.5	— 19.6	— 24.3
	ΔI_a in 10^{36} g cm^2		— 2.00		— 1.87	— 0.57	
	Both hemispheres ΔI_a		0.81		— 0.83	0.06 — 0.08 (equ. 6) — 0.02	

where the subscripts refer to integration over the entire surface of the earth, over the ocean areas, and over the land areas, respectively. The ocean water will move in such a manner that Δm_o has the same value over the entire ocean area. It follows that

$$\int\int_o \Delta m_o \, ds = \Delta m_o \int\int_o ds = \Delta m_o \, S_o$$

$$= -\int\int_L \Delta m_L \, ds = -\overline{\Delta m_L} \int\int_L ds = -\overline{\Delta m_L} \, S_L$$

or

$$\Delta m_o = -\frac{S_L}{S_o} \overline{\Delta m_L}, \qquad (5)$$

where $\overline{\Delta m_L}$ represents a mean value of Δm_L over that portion of the earth's surface which is covered by land.

Changes in the moment of inertia are composed of those occurring over the ocean and those occurring over land:

$$\Delta I_a = R^2 \Delta m_o \int\int_o \cos^2 \varphi \, ds +$$

$$+ R^2 \int\int_L \Delta m_L \cos^2 \varphi \, ds =$$

$$= - R^2 \frac{S_L}{S_o} \frac{\overline{\Delta p_L}}{g} \int\int_o \cos^2 \varphi \, ds +$$

$$+ \frac{R^4}{g} \int_{-\pi}^{\pi} \int_L \Delta p_L \cos^3 \varphi \, d\varphi \, d\Theta, \qquad (6)$$

where $\Delta p = g \Delta m$, and

$$\overline{\Delta p_L} = g \overline{\Delta m_L} = \frac{R^2}{S_L} \int_{-\pi}^{\pi} \int_L \Delta p_L \cos \varphi \, d\varphi \, d\Theta \qquad (7)$$

is the mean (local) pressure variation over all land. The computation is summarized in Table 2. For the earth, $S_L/S_o = 0.419$, $S_E = S_L + S_o = 5.10 \times 10^{18}$ cm^2, $\int\int_o \cos^2 \varphi \, ds = 0.49 \, S_E$, $\overline{\Delta P_L} = 0.183$ mb $= 183$ dynes cm^{-2}, and the first term in (6) reduces to $-4.32 \times 10^{32} \, \overline{\Delta P_L} = -0.08 \times 10^{36}$ g cm^2, compared to $+0.06 \times 10^{36}$ g cm^2 for the second term. In view of the uncertainty of the calculations, this high degree of cancellation is largely fortuitous, and the final value of ΔI_a may be as large as 0.25×10^{36} g cm^2 and of either sign. We shall adopt this larger value. For comparison, the moment of inertia of a one meter snow cover, or 20 cm water cover, over the entire United States is also 0.25×10^{36} g cm^2. It would be desirable to recompute Table 2 using the original data for the uncorrected local pressure.

A larger effect is related to the shifting of ocean currents, and the distribution of mass associated with the currents, by the variable winds. This effect could best be computed from seasonal changes in the topography of the sea surface as obtained either from standard oceanographic observations of density versus depth, or from tide gauges. Unfortunately there are very few ocean areas from which repeated oceanographic observations have shed light on seasonal variations. Many tide gauge records show seasonal variations in sea level (Δz) of the order of 20 cm, but almost all these variations are associated with coastal currents and other local phenomena.

In order to obtain a rough estimate of the sea level effect on a world-wide basis we have noted a relationship (MUNK, 1950) between the mean annual sea level in the Atlantic from 50° N to 60° S, and the corresponding distribution of atmospheric pressure over the same area. Both the sea level and the pressure reach maxima in 30° N and 30° S and drop sharply towards higher latitudes. The "dynamic sea level" is therefore opposite in phase from the "hydrostatic sea level" previously considered. To obtain a numerical relationship we must first subtract the hydrostatic effect and write

$$\frac{\varrho_w \, g \, z_w + p_{\text{atm}}}{p_{\text{atm}}} = \lambda,$$

where ϱ_w is the density of water, z_w the deviation of sea level from its mean, and λ a proportionality factor which varies from 2.5 near the equator to about 7.5 at 60°, and can be represented by $\lambda = 2.5 + 5 \varphi^2$, where φ is the latitude in radians.

We now assume that this relationship $\lambda(\varphi)$ can be applied to seasonal variations over all ocean areas, and thus compute the dynamic sea level effect from the known seasonal pressure variations according to

$$\Delta I_o = \frac{R^4}{g} \int_{-\pi}^{\pi} \int_o \lambda (\Delta p_o - \overline{\Delta p_o}) \cos^3 \varphi \, d\varphi \, d\Theta, \qquad (8)$$

where $\overline{\Delta p_o} = -(S_L/S_o) \, \overline{\Delta p_L} = -0.77$ mb is the mean pressure variation over all oceans. At any latitude

$$\int_o \lambda \Delta p_o \cos^3 \varphi \, d\Theta = \lambda \left\{ \int_o^{2\pi} \Delta p \cos^3 \varphi \, d\Theta - \int_L \Delta p \cos^3 \varphi \, d\Theta \right\},$$

and the first term in (8) can be obtained by subtracting Column 6 from Column 5 in Table 2, and multiplying by appropriate values of λ. The second term,

$$- \frac{R^4}{g} \overline{\Delta p_o} \int\int_o \lambda(\varphi) \cos^3 \varphi \, d\varphi \, d\Theta$$

has been evaluated numerically and equals 1.3×10^{36} g cm^2 over the northern hemisphere and 1.8×10^{36} g cm^2 over the southern hemisphere. The results are

ΔI_o, northern hemisphere 1.4×10^{36} g cm^2
ΔI_o, southern hemisphere -2.2×10^{36} g cm^2
ΔI_o, both hemispheres -0.8×10^{36} g cm^2.

The extrema in pressure changes and in the computed dynamic sea level changes are summarized in Table 3.

Table 3. Extrema in pressure changes ($\Delta p_o - \overline{\Delta p_o}$) and sea level changes Δz

Latitude	λ	$\Delta p - \overline{\Delta p_o}$ mb	Δz cm
67° N	9	+ 3	+ 27
57° N	7	— 5	— 35
20° N	3	+ 4	+ 12
20° S	3	— 3	— 9
50° S	6	+ 3	+ 18
70° S	10	— 3	— 30

The computed changes in sea level are of the same magnitude as those obtained from tide gauge records.

There remain the changes ΔI_e in the inertia of the solid earth and biosphere. To the first of these have been ascribed rather abrupt changes in $\Delta\Omega_e/\Omega_e$ as large as 4.0×10^{-8}, which appear to occur every two or three decades. Whatever the cause of these intermittent changes, it is unlikely that *seasonal* changes in the moment of inertia of the solid earth are of sufficient magnitude to compare with the effects discussed above[5]. The effects of vegetation, related to the rise of sap from the ground into the woody part of trees, of the production of leaves, and of other factors, have been estimated by JEFFREYS (1929) to have a seasonal range of 2 g cm^{-2} in the temperate zones, corresponding to a change in the moment of inertia by 0.15×10^{36} g cm^2 for the northern hemisphere. Compared to these values, any direct effect of human activity is negligible. If all American automobiles were driven from Alaska to Mexico, the change in the moment of inertia would amount to only 10^{31} g cm^2.

The combined changes of inertia of atmosphere, ocean and earth probably do not exceed 1.00×10^{36} g cm^2. This leads to a value of $\Delta I/I \cdots 0.12 \times 10^{-8}$.

[5] Dr Mintz has pointed out the interesting possibility that seasonal variations in the solar tides may have an important effect.

Summary of seasonal fluctuations

The interrelationships between the various atmospheric and oceanic terms appearing on the right side of equation (2) is sketched in fig. 1. The magnitudes are as follows:

$\Delta(I_a \omega_a)/I\Omega_e$ ..... $+ 1.6 \times 10^{-8}$
$\Delta(I_o \omega_o)/I\Omega_e$ ..... 0.15×10^{-8} (plus or minus)
$\Delta I/I$ ..... 0.12×10^{-8} (plus or minus)

The term on the left side of equation (2), $-\Delta\Omega_e/\Omega_e$, has been determined by STOYKO (1936, 1937, 1949) and FINCH (1950) from astronomic observations:

STOYKO
1934—1935 ..... 2.93×10^{-8}
1935—1936 ..... 0.76×10^{-8}
1936—1937 ..... 1.59×10^{-8}
1946—1947 ..... 2.40×10^{-8}

FINCH
1943—1947 ..... 2.0×10^{-8}
1948—1949 ..... 2.3×10^{-8}

Here $\Delta\Omega_e$ represents January rotation minus the preceeding July rotation.

The observations for 1934—1935 can be compared to those by SCHEIBE and ADELSBERGER (1936). These give 1.4×10^{-8} or

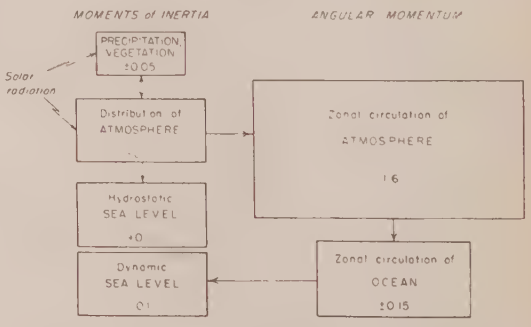

Fig. 1. Schematic presentation of the relationship between factors influencing the earth's angular velocity. The numbers give an indication of the seasonal effect of each factor, expressed in parts per hundred million of the mean rate of rotation, or roughly in milliseconds per day. (+) represents excess of earth's rotation in July over that in January. (±) represents magnitude without indication of the direction of the effect. Sea level changes are composed of those induced by the hydrostatic pressure changes, and those related to the oceanic circulation, which in turn is caused by the winds.

2.5×10^{-8}, depending on whether the difference January minus July or January minus August are formed. For the preceeding year the seasonal variation appears to be small and obscured by the scatter of data.

The meteorologic and astronomic observations agree with regard to both magnitude and sign. The astronomic observations reveal a difference in rotation between March and August which is, on the average, 80 per cent larger than the difference between January and July. The meteorologic implication of this observation is not at all in disagreement with synoptic experience.

Accuracy of determining earth's angular velocity

The detection of short period anomalies in the atmospheric circulation is likely to be of greater practical interest in meteorology than the seasonal fluctuations. Whether the earth's rotation can be determined with sufficient accuracy to make this possible depends largely on two factors: the accuracy of time determination from the zenith passage of stars, and the precision of the clocks with which the astronomic time is to be compared.

By using a "bank" of *quartz crystal clocks*, and "beating" the clocks against one another, it has been possible to achieve a relative accuracy of one part in a hundred million. The term relative is used, because it is found that the frequency characteristics of the crystals undergo sudden small changes from time to time, and the clocks must be compared to some independent time standard, i.e., the mean annual rate of the earth's rotation. In 1898, and again in 1920, the mean annual rate of rotation changed rather abruptly, and during those years a determination of the seasonal components of the earth's angular velocity would have led to erroneous results.

The development of *atomic* clocks promises to provide an absolute time standard independent of the earth's rotation. In the case of the *ammonia* clock now being developed by LYONS (1949) a constancy of one part in 20 million has been achieved, and an accuracy of one part in 100 million is hoped for. However, in another development carried on by the Bureau of Standards with the assistance of Professor P. KUSCH of Columbia University, quantum transitions in beams of atoms such as *cesium* will be used to establish frequency and time standards. The potential accuracy is increased (Bureau of Standards, 1949) "by a factor of 10 to 100 or more", and "an ultimate accuracy of 1 part in 10 billion may be reached."

A quite different limitation is imposed by the accuracy to which the "earth clock" and the atomic clock can be *read* in order to measure the cumulative difference between the rates of the two clocks. In practice only the accuracy in reading the earth clock need be considered. The present accuracy of astronomic time determination is about 0.02 seconds. According to the Astronomer Royal (SPENCER JONES, 1945) the proposed installation at the Royal Observatory of 18 quartz crystal standard clocks in conjunction with the photographic zenith tube will reduce the probable error in time determination to 0.002 or 0.003 seconds. Dr. SOLLENBERGER at the U. S. Naval Observatory estimates (personal communication) that the error can ultimately be reduced to 0.001 or 0.002 seconds.

We may interpret these values in terms of the probable accuracy to which the earth's angular velocity can be determined. Suppose we wish to measure the mean angular velocity of the earth during a period of N days, starting with $t = -1/2\,N$ and ending with $t = +1/2\,N$. Once each day we determine the difference x between the astronomic time y and the clock time z. The slope m of the line $x = f(t)$ equals

$$m = \frac{\Delta \Omega_e}{\Omega_e} = \frac{\sum_{t=-N/2}^{N/2} x_i\, t_i}{\sum_{t=-N/2}^{N/2} t_i^2}$$

by the methods of least squares, with the understanding that the mean value of x has been subtracted from the individual values.

The probable error m' of m equals the geometric sum of the probable errors of each term in the series:

$$m'^2 = \frac{\sum_{t=-N/2}^{N/2} (x_i'\, t_i)^2}{\left(\sum_{t=-N/2}^{N/2} t_i^2\right)^2}. \qquad (9)$$

The probable error x_i' of $x_i = y_i - z_i$ equals

the geometric sum of the probable errors of y_i and z_i:

$$x_i'^2 = y_i'^2 + z_i'^2.$$

Here y_i' is the error in astronomic time determination, which remains constant throughout the time interval N. The probable clock error may be set as $z_i' = \omega' t_i$, where ω' is the constant probable error in the *rate* of the clock relative to its mean rate. Substituting

$$x_i'^2 = y'^2 + \omega'^2 t_i^2$$

into Equation (9) leads to

$$2 m'^2 = \frac{y'^2}{F} + \frac{f\omega'^2}{F^2}, \qquad (10)$$

where

$$F(N) = 1^2 + 2^2 + \ldots (N/2)^2 = \frac{N^3}{24} + \frac{N^2}{8} + \frac{N}{12},$$

$$f(N) = 1^4 + 2^4 + \ldots (N/2)^4 = \frac{N^5}{160} + \frac{N^4}{32} + \frac{N^3}{24} - \frac{N}{60}.$$

Both N and y' are given in days.

The uncertainties in the astronomic time determination and in the rate of the clock contribute equally to the uncertainty in the determination of the earth's rotation if

$$\frac{y'^2}{\omega'^2} = \frac{f}{F} \approx \frac{3N^2}{20}.$$

The present accuracy of $y' = 0.02$ seconds $\approx 2 \times 10^{-7}$ days, $\omega' = 10^{-8}$, gives $N \approx 52$ days. For periods much longer than 2 months the error is due chiefly to the error in the rate of the clock. For much shorter periods the error is due chiefly to the error in astronomic time determination, and equals

$$m' = \frac{y'}{\sqrt{2F}} \approx \frac{\sqrt{12}\, y'}{N^{3/2}},$$

where the approximation holds for all but very small values of N. The chief limitation at present is imposed by the astronomic time determination rather than by the clock.

Conclusions

In fig. 2 the probable error in the determination of the anomalies in the earth's rotation has been plotted against the length of time over which the average has been taken. The lines correspond to the four possible combinations between astronomic time determination (p.e. of present method 0.02 sec, of photographic zenith telescope 0.002 sec) and standard time determination (p.e. of crystal clocks $1 : 10^8$, of cesium clock $1 : 10^9$). According to fig. 2 the probable error of monthly averages now is 5×10^{-9}. The probable error of Stoyko's measurements, judged externally from the consistency of his observations, is somewhat smaller, and our values for the accuracy of astronomic time determination and the zenith tube can be regarded as conservative.

The shaded band in fig. 2 is an estimate of the desirable accuracy. For example, the measurement of seasonal variations of two

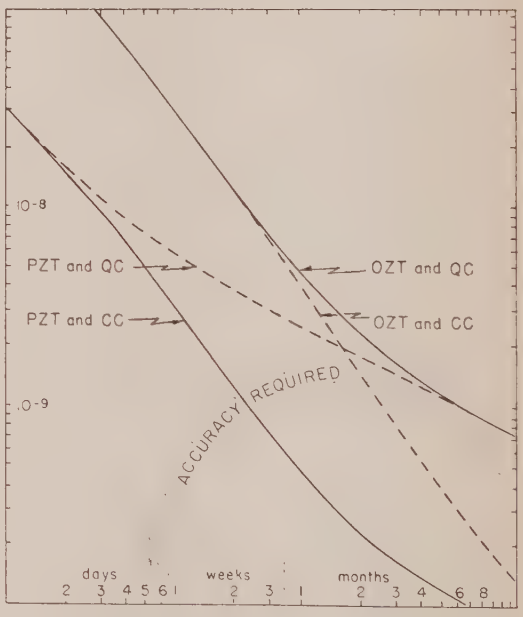

Fig. 2. Probable fractional errors in determining the earth's angular velocity as a function of the length of time over which the measurement of the velocity are averaged. The four lines correspond to the possible combinations of the ordinary zenith tube now used (OZT, $\pm$ .02 sec.), and the proposed photographic zenith telescope (PZT, $\pm$ .002 sec.), with the quartz clock (QC, $\pm 10^{-8}$), and the proposed cesium clock (CC, $\pm 10^{-9}$). The shaded band is an estimate of the largest probable error permissible for determining fluctuations in the earth's rotation of given duration.

parts per hundred million requires monthly (or at least quarterly) values accurate to, say, two parts per billion. The present accuracy is therefore too small for a reliable measurement of these variations, but not too small for their detection.

In order to study fluctuations of annual means insofar as they are related to meteorological conditions it will be necessary to improve the accuracy of the clock, and furthermore to refer to an absolute time standard. Both requirements are fulfilled by the cesium clock.

Anomalies persisting for short periods of time will be relatively small, and the required accuracy correspondingly high. Since the "half life" of atmospheric circulation is only a week we may expect anomalies persisting for a week to be as large as perhaps 25 per cent of the monthly anomalies. The required accuracy can then be set at one part in two billion. According to fig. 2 the development of the photographic zenith tube and of the cesium clock should enable us to measure anomalies persisting for two weeks or more. By relaxing our requirements and assumptions these values can be improved by a factor of two. Any further improvement will depend largely upon our ability to correct the astronomic time determination during zenith passage for the refraction of light from a distant star by the atmosphere directly over the telescope. Thus the ultimate limitation on measuring atmospheric circulation by this method is imposed by the atmosphere itself. It is hoped that the great interest attached to an index of *integrated* circulation as derived from anomalies in the earth's rotation might provide an additional incentive to improve present methods of time determination.

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Ueber die allgemeine atmosphärische Zirkulation während der diluvialen Vereisungsperioden

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Abstract

The range of polar cold air over the northern hemisphere was increased during the glacial periods primarily due to the extension of the large ice shields. Particulary in the northatlantic region the atmospheric centers of action and the storm tracks were shifted southwards, the meridional temperature gradient was strengthened and consequently the general circulation increased. This intensification of the large scale exchange of air does not contradict the recent increase of the general circulation and its consequences, since during the glacial periods the increase was of a compulsory and secondary nature caused by the ice extension itself. On the contrary the atmospheric circulation at the beginning of the glaciation must have been decreased. However, the continued extension of the Scandinavian ice shields, as well as the creation of the Keewatin and Labrador shields in North America and the North Asian glaciation, occurred during periods of secondarily increased circulation. An examination of the most important Ice age theories shows that the primary weakening of the circulation may be satisfactorily explained only by assuming a decrease of the earth's solar radiation supply. The astronomical Ice age theory is applicable only in connection with the assumption of an uplift of the Iceland-Faerisland submarine ridge.

In den letzten 20 Jahren sind unsere Kenntnisse der klimatischen Verhältnisse im Diluvium durch zahlreiche Arbeiten von geologischer, meteorologischer, geographisch-morphologischer und botanischer Seite ausserordentlich erweitert worden. Es liegt daher der Gedanke nahe, von der Analyse der regionalen Klimate zu einer Synthese des Zustandes der Atmosphäre überzugehen und zu versuchen, Ablauf und Intensität der allgemeinen Zirkulation in den Vereisungsperioden zu rekonstruieren. Es finden sich zwar in mehreren Arbeiten über das Diluvialklima Hinweise auf die Zirkulationsverhältnisse — besonders bei KLUTE (1930), WUNDT (1943), MEINARDUS (1937) und BÜDEL (1949) —, jedoch fehlt bisher eine Koordinierung der aus den Beobachtungen abgeleiteten Schlussfolgerungen mit den für Entstehung, Erhaltung und Vergehen des Inlandeises zu fordernden Verhältnissen.

Die mangelnde quantitative Erfassung und Deutung des gegenwärtigen Luftkreislaufes und das Fehlen von genaueren Angaben über die meteorologischen Elemente und Faktoren in der geologischen Vorzeit verhindern eine exakte Behandlung der Paläozirkulation. Man muss sich daher auf quantitative Betrachtungen beschränken und die jeweilige Zirkulation als einen vom Temperatur- und Druckgradienten abhängigen Austauschvorgang mit den gegenwärtigen Verhältnissen vergleichen. Voraussetzung für eine derartige Behandlungsweise ist, dass das Grundschema des Luftkreislaufes damals das gleiche war wie heute. Diese Bedingung dürfte ohne weiteres erfüllt sein, da keine Anzeichen und Gründe für Veränderungen der den atmosphärischen Vor-

gängen zugrundeliegenden physikalischen Gesetze vorhanden sind.

1. Diluvialklima und Zirkulation

Es gilt zunächst zu untersuchen, welche Aussagen über die allgemeine Zirkulation auf Grund der »beobachteten« Klimaverhältnisse zur Zeit des Höchststandes der letzten Vereisung — bezw. während einer Zeit maximaler Eisausdehnung — gemacht werden können. Vergleicht man die gegenwärtige Eisbedeckung der polaren Gebiete und die Verteilung der Klimate mit den Verhältnissen während der Glazialzeiten, so ergibt sich, dass auf der Nordhemisphäre durch die sich in Mitteleuropa bis etwa 55° N, in Nordasien bis etwa 60—65° N und in Nordamerika sogar teilweise bis zum 40. Breitengrad erstreckenden Eismassen die Grenze des polaren (bezw. nivalen) Klimagürtels bis zu 25 Breitengrade südwärts verschoben war. Dementsprechend mussten sich auch die übrigen Klimabereiche äquatorwärts verlagern, worauf erstmalig Penck (1913) hingewiesen hat und was später von Klute (1930) und kürzlich von Büdel (1949) durch die Verschiebung der Vegetations- und Landschaftsgürtel im mittel- und südeuropäischen Raum anschaulich dargestellt worden ist. Auf der Südhemisphäre muss eine gleichsinnige Verschiebung der Klimazonen stattgefunden haben, allerdings in weitaus schwächerem Umfange, da die antarktische Eiskappe gegenüber der Jetztzeit nur eine geringe Zunahme der Flächenausdehnung (bis zum Schelfrand) erfahren haben kann. Gleichzeitig mit der Verschiebung der Klimazonen bezw. in kausalem Zusammenhang damit müssen naturgemäss auch Änderungen in den grossräumigen Strömungsverhältnissen stattgefunden haben, d. h. die atmosphärischen Aktionszentren müssen gegenüber ihrer jetzigen Lage verschoben gewesen sein.

Für den nordatlantisch-europäischen Raum ergibt sich folgendes Bild:

1. Die nordatlantische Depression (Islandtief) war durch die ausgedehntere Eisbedeckung des Skandik — Scholleneis, Eisberge und mindestens im Winter auch Packeis — südwärts auf etwa 50° Nordbreite (mit jahreszeitlichen Schwankungen von etwa 10—15 Breitengraden) verlagert und ihre Intensität infolge der stärkeren Temperaturgegensätze zwischen dem relativ warmen Nordatlantik (Golfstrom) und dem europäischen Nordmeer gegenüber der Jetztzeit verstärkt.

   Die Annahme von Penck (1937), dass der Golfstrom damals nicht vorhanden war, trifft sicher nicht zu. Es ist zwar wahrscheinlich, dass seine Intensität infolge der veränderten Küstenkonfiguration im westindischen Raum (eustatische Senkung des Meeresspiegels) damals schwächer war; er muss jedoch als Teil der nordatlantischen Zirkulation unbedingt vorhanden gewesen sein. Zudem entfällt bei Annahme eines völlig kalten Nordatlantik jede Möglichkeit, die Herkunft der zur Bildung des eurasischen Inlandeises notwendigen Feuchtigkeitsmengen zu erklären (Behrmann 1948).
2. Über die Lage der nordatlantischen Hochdruckzelle sind keine genaueren Aussagen möglich. Es kann angenommen werden, dass auch dieses Aktionszentrum im Jahresmittel um einige Breitengrade südwärts verlagert war.
3. Die Temperaturgegensätze in meridionaler Richtung zwischen den gemässigten und subtropischen Breiten sowie zonal zwischen den eisfreien Meeres- und Landgebieten der mittleren Breiten waren erheblich grösser als gegenwärtig. Das Quellgebiet der Arktikluft reichte damals entsprechend der Eisbedeckung 20—25 Breitengrade weiter südwärts als heute.

   Der häufig vorgetragene Einwand, dass eine Verschärfung der meridionalen Temperaturgegensätze in den Glazialperioden nicht anzunehmen sei, da es Zeiten einer generellen Temperaturabnahme auf der Erde waren, ist nicht stichhaltig. Es werden zwar bei einer Verminderung der solaren Zustrahlung — sofern man das als Ursache der diluvialen Klimaänderungen ansieht — die tropischen Gebiete stärker betroffen als die höheren Breiten (Ausgleich der Gegensätze), jedoch muss in den letzteren Gebieten als sekundärer Effekt die verstärkte Strahlungsreflexion (Wundt 1938 a) und der Verbrauch von Schmelzwärme durch die Eismassen berücksichtigt werden. Für die Verstärkung der

Gegensätze zwischen Land und Meer wird besonders die sommerliche Erwärmung der periglazialen Gebiete gegenüber dem durch die Eisschmelze kühlen Nordatlantik ausschlaggebend gewesen sein.

4. Aus der Verlagerung des nordatlantischen Zentraltiefs und der Verstärkung der meridionalen Temperaturgegensätze ergibt sich zwangsläufig, dass in den Glazialperioden der meridionale Druckgradient Azorenhoch — Islandtief verschärft und damit der Luftaustausch zwischen den mittleren und subtropischen Breiten, d. h. die aussertropische Zirkulation, verstärkt waren.
5. Aerologische Untersuchungen über die jetzige Höhenlage der Tropopause legen den Schluss nahe, dass in den Glazialzeiten das meridionale Gefälle dieser Diskontinuitätsfläche erheblich stärker war als heute. Entsprechend der Ausdehnung der Arktikluft werden damals im mitteleuropäischen Raum ähnliche oder noch extremere Verhältnisse geherrscht haben wie heute im Winter Nordostsibiriens, wo die Tropopause mit weniger als 8 km Höhe ihre niedrigste Lage erreicht. — Die von KLUTE (1930) angenommene Absenkung der Tropopause über dem Inlandeis bis auf 1 km Höhe dürfte jedoch unzutreffend sein. — Daraus ergibt sich, dass wahrscheinlich auch an der Tropopause der Druck- und Temperaturgradient verstärkt waren.
6. Infolge der Verschärfung der meridionalen Druck- und Temperaturgegensätze war während der Glazialzeiten die atlantische Polarfront sehr kräftig entwickelt (zeitweiliger Zusammenfall von Arktik- und Polarfront in heutigem Sinn) und wegen der Ausdehnung der Arktikluft südwärts verschoben. Dementsprechend müssen auch die sich an der Front ausbildenden Wellenstörungen sehr intensiv und von längerer Lebensdauer als heute gewesen sein und sich zudem auf im Mittel südlicheren Bahnen bewegt haben.

Im Sommer verlief die Frontalzone wahrscheinlich über Mitteleuropa, und die wegen der starken Temperaturgegensätze zwischen den periglazialen und vereisten Gebieten ausserordentlich kräftig ausgebildeten und zahlreichen Depressionen drangen sowohl auf der Zugstrasse I über die Eisgrenze weit nach Norden und Nordosten als auch auf den südlicheren Zugstrassen (IV und V) weit ostwärts vor. Die Annahme einer als Sperre wirkenden hochreichenden glazialen Antizyklone über dem nordischen Inlandeis (im Sinne von HOBBS) kann wohl nach den neueren Untersuchungen über die grönländischen Verhältnisse (GEORGI 1939, DORSEY 1945) endgültig verlassen werden. Die mit der planetarischen Westdrift vom Nordatlantik heraufziehenden Zyklonen rollten entweder die relativ flache, bodennahe Kaltluft auf oder wurden durch diese vom Boden abgehoben, okkludierten und drangen als Warmluftreste weiter »eiseinwärts» vor (BLÜTHGEN 1942). Nur auf diese Weise konnten den Eismassen die für ihre Ernährung und Erhaltung notwendigen Niederschlagsmengen zugeführt werden. — Im Winter wird sich die Hauptzyklonentätigkeit entsprechend der etwas südlicheren Lage der Frontalzone vorwiegend auf den südlichen Zugstrassen abgespielt haben.[1] Für den mediterranen Raum ergibt sich damit: im Jahresmittel Vorherrschaft westlicher Winde (stärkere Senkung der eiszeitlichen Schneegrenze an den Westhängen der mediterranen Gebirge gegenüber den Osthängen!) und gegen heute eine Zunahme der jährlichen Niederschlagsmenge, womit die im Mittelmeerraum und in Nordafrika beobachteten »Pluvialzeiten», d. h. die Südwärtsverlagerung vom polaren Rand des nördlichen Trokkengürtels, eine zwanglose Erklärung finden (KLUTE 1930).

Zusammenfassend ist festzustellen, dass während der Glazialzeiten im nordatlantisch-europäischen Raum die aussertropische Zirkulation gegenüber der Jetztzeit allein auf Grund der Anwesenheit der nordischen Eisbedeckung verstärkt gewesen sein muss. — Es sei bemerkt, dass BÜDEL (1949) auf Grund der Verbreitung

[1] Nach Fertigstellung des Manuskriptes erhielt ich von Herrn Prof. Dr. F. KLUTE freundlicherweise einen Sonderdruck seiner letzten Veröffentlichung über das Diluvialklima zugesandt (KLUTE 1949). Die in vorliegender Arbeit skizzierten Zugstrassen der Depressionen sind u. a. darin in ihrer Bedeutung näher erläutert und in Kartenskizzen anschaulich dargestellt.

der Lösstundra und Löss-Steppe in Europa für eine Abschwächung der Westdrift eintritt, worauf an anderer Stelle näher eingegangen werden soll.

Eine analoge Ableitung der atmosphärischen Verhältnisse für die übrigen Gebiete der mittleren Nordbreiten ist wegen mangelnder Beobachtungsgrundlagen (mit Ausnahme des nordamerikanischen Raumes) nicht möglich. Die allgemeinen Grundtatsachen der Diluvialmeteorologie deuten jedoch durchweg auf eine Verstärkung des aussertropischen Luftkreislaufes während der Vereisungsperioden:

1. In Nordamerika herrschten bezüglich der Ausbreitung der Arktikluft und der Verschiebung der Klimazonen ähnliche Verhältnisse wie in Europa, d. h. der meridionale Temperaturgradient war auch hier erheblich verschärft. Dadurch wurden die Zyklonentätigkeit und die Häufigkeit der in diesem Gebiet gefürchteten Kaltlufteinbrüche und der Warmluftvorstösse vergrössert und ihre Intensität verstärkt. Da die Kaltlufteinbrüche Nordamerikas (Blizzard) in der Gegenwart allgemein weitaus stärkere Schneefälle auslösen als die des mitteleuropäischen Raum, kann die erwähnte Intensivierung der meridionalen Temperaturunterschiede in den Vereisungsperioden und der sich daraus ergebende in Nordamerika völlig ungehinderte meridionale Luftaustausch als ein wesentlicher Grund für die grössere Mächtigkeit des nordamerikanischen Inlandeises gegenüber dem europäischen angesehen werden.
2. Die diluvialen Eisverhältnisse im nordpazifischen Raum sind weitgehend unbekannt. Es ist wahrscheinlich, dass mindestens zur Zeit maximaler Eisausdehnung (Riss) die bis etwa 62° N reichende Landbrücke zwischen Asien und Nordamerika (eustatische Senkung des Meeresspiegels — heutige Meerestiefe bis 62° N maximal 66 m) vereist und weitere Gebiete der Beringsee packeisbedeckt waren. Je nach Stärke der Eisbedeckung in diesen Gebieten muss das Aleutentief mehr oder weniger südwärts verlagert und damit der Druckgradient in den mittleren Breiten des Nordpazifik verschärft gewesen sein.
3. Die stärkere Vereisung des Nordwest-Pamir im Diluvium ist, wie schon v. Ficker (1933) feststellte, nur durch eine Vermehrung der Niederschläge, d. h. durch eine Verstärkung der Zyklonentätigkeit in diesem Gebiet zu erklären.
4. Das Gleiche gilt auch für die nord- und nordostasiatischen Eisschilde, die sich in Gebieten entwickelten, welche heute zum polaren Trockenklima gehören.

In diesem Zusammenhang sei auf die Untersuchung von Albrecht (1947) hingewiesen, in der gezeigt wird, dass gegenwärtig der gesamte asiatische Raum nordwärts des 50. Parallel ausschliesslich aus dem nordatlantischen Verdunstungsgebiet mit Feuchtigkeit versorgt wird, da die küstenparallelen Gebirge Ostsibiriens das Eindringen feuchter Strömungen vom Pazifik verhindern. Die nordsibirischen Eisgebiete können somit nur durch eine verstärkte Wasserdampfzufuhr vom Nordatlantik bzw. durch häufiges Vordringen zyklonaler Störungen etwa auf der Bahn Mittelmeer-Schwarzmeer-Kaspisee nach Nordosten erklärt werden.

Auf der Südhemisphäre sind wegen der geringen Landbedeckung in den gemässigten Breiten nur sehr wenige Ansatzpunkte vorhanden, die Rückschlüsse auf die diluvialen Zirkulationsverhältnisse zulassen. Der antarktische Kontinent befindet sich gegenwärtig zweifellos noch immer in einer Glazialperiode, jedoch würden an zahlreichen Stellen Antarktikas Anzeichen einer noch stärkeren Vereisung in jüngerer Vergangenheit beobachtet. Diese verstärkte Eisbedeckung konnte sich aber weniger in einer vergrösserten Flächenausdehnung als in einer Zunahme der Mächtigkeit des Eises auswirken, da eine Ausbreitung der Eiskappe nur bis zu dem vom heutigen Eisrand allgemein nicht weit entfernten Schelfrand möglich war. Jedoch wird die Packeisbedeckung der südpolaren Meeresgebiete sicher ausgedehnter gewesen sein als gegenwärtig. Daraus folgt, dass der für die Nordhalbkugel abgeleitete Effekt: Ausdehnung der Eisgebiete — Verschiebung der Klimazonen und Aktionszentren — Verstärkung der aussertropischen Zirkulation in den mittleren und hohen Südbreiten sicher erheblich schwächer ausgeprägt war. Es liegen aber auch aus Chile, Südafrika und Südaustralien Beobachtungen vor (Klute 1930), dass hier in jüngster Vergangenheit — analog den nordhemisphärischen Verhältnissen — der polare Rand des südlichen Trockengürtels

äquatorwärts verschoben war. Man kann somit feststellen, dass auch auf der Südhalbkugel in den Glazialperioden des Diluviums die aussertropische Zirkulation gegenüber der Jetztzeit verstärkt war.

Derartige Verhältnisse werden von Brüggen und Mortensen auch zur Erklärung der starken Depression der eiszeitlichen Schneegrenze in den nord- und mittelchilenischen Anden sowie der mächtigen Talschotter in den gegenwärtig vollariden Gebieten Nordchiles gefordert. Dabei kann angenommen werden, dass der sogen. abnorme Wettertyp Chiles mit kurzen aber ausserordentlich kräftigen Niederschlägen gegenüber heute erheblich öfter und anhaltender eingetreten ist (Viete 1949).

Nach der Besprechung der aussertropischen Zirkulation beider Hemisphären gilt es nun noch den tropischen oder Passatkreislauf in den Glazialperioden zu untersuchen.

1. Im äquatorialen Afrika konnten Nilsson (1940) im Hochland von Semien (Abessinien), am Ruwenzori, Mt. Kenya und Mt. Elgon sowie H. Meyer und Klute am Kilimandscharo starke mehrmalige Senkungen der Schneegrenze im Diluvium und z. T. 2—3 Vergletscherungsphasen nachweisen. Aus der Hangdisposition der eiszeitlichen Gletscherkappe des Kilimandscharo im Vergleich zur rezenten schloss Klute, dass damals wie heute die niederschlagsbringenden Winde aus östlicher Richtung kamen.
2. Durch eingehende Untersuchungen von Strandlinien und Terrassen an Seen in Britisch-Ostafrika gelang Nilsson (1940) der Nachweis von 3 Pluvialperioden, von denen die zweite am kräftigsten ausgeprägt war und die sich eindeutig mit den erwähnten Vergletscherungsphasen parallelisieren lassen.
3. In Indien wiesen De Terra und Paterson (1939) im Gebiet von Kaschmir 4 Glazialperioden und in Nordwest-Punjab sowie in den Tälern des Indus und Chenab und in Zentralindien 4 diesen Eiszeiten parallele Pluvialperioden nach. Aus der Ablagerung des diluvialen Löss in Kaschmir und in Nordwest-Punjab folgerten sie, dass damals die mittlere Windrichtung zur Zeit der Sedimentation die gleiche war wie heute, dass aber die Intensität der gegen den Himalaya gerichteten Luftströmungen erheblich stärker gewesen sein muss.
4. Von wesentlicher Bedeutung ist auch das Problem der Verschiebung von den äquatorialen Rändern der subtropischen Trockenzonen. Während Penck (1913) zunächst eine allgemeine Verschiebung der ariden Gebiete zum Äquator, d. h. eine Einengung der humiden Tropenzone feststellte, ergaben die Untersuchungen von Wissmann in China, Jaeger in Mexiko, Troll in den bolivianischen Anden u. a. sowie die Arbeiten von Klute (1930) und Wundt (1938), dass in der Eiszeit »nicht die regenreichen Tropen eingeengt waren, sondern die Wüstengürtel an ihren Flanken» (Penck, 1936).

Eine jeder Kritik standhaltende Deutung dieser Beobachtungen aus dem Tropengürtel bezüglich der Zirkulationsverhältnisse ist leider nur schwer zu geben. Zunächst kann wohl als gesichert angesehen werden, dass die tropischen Pluvialperioden nicht allein durch eine Verringerung der Verdunstung infolge der allgemeinen Temperaturabnahme, sondern auch durch eine Vermehrung der Niederschläge zu erklären sind. In Ostafrika hängen die Niederschlagsverhältnisse in hohem Masse von der Lage des Druckmaximums im Südindik und von der Intensität des SE-Passates im Südwinter ab. Durch eine Verstärkung dieser Strömung in den Glazialzeiten infolge (geringer) Nordwärtsverlagerung der im Südindik gelegenen Hochdruckzelle und der — wohl in Übereinstimmung mit Wundt (1938) als Folge der ausgedehnten nordhemisphärischen Eisbedeckung als gesichert zu betrachtenden — südlicheren Lage des thermischen Äquators können die sommerlichen Niederschläge im ostafrikanischen Raum erhöht worden sein, während im Winter die allgemeine Verminderung der Verdunstung ihren Teil zur Erhaltung der hohen Wasserstände der dortigen Seen beitrug. – Die Beobachtungen aus Indien deuten zweifellos auf eine Verstärkung des Sommermonsuns, deren Ursache möglicherweise ebenfalls in der erwähnten Intensivierung des SE-Passates, der im Bereich des Äquators nach NE umbiegt, gesucht werden kann. Andererseits muss aber auch ein sehr kräftiger Wintermonsun in diesen Gebieten vorhanden gewesen sein,

da in den Glazialepochen das sibirische Kältemaximum zweifellos kräftiger ausgebildet war als heute (VIETE 1949).

Diese kurzen Ausführungen zeigen bereits, dass genauere Aussagen über die Strömungsverhältnisse in den niederen Breiten erst nach einer wesentlichen Erweiterung des Beobachtungsmaterials möglich sind. Sie machen wahrscheinlich, dass im Bereich des nördlichen Indik sowohl die Monsunzirkulation als auch der planetarische Kreislauf verstärkt waren. Für eine allgemein grössere Intensität der tropischen Zirkulation als beste Verwirklichung einer thermisch angeregten Meridionalzirkulation sprechen ausserdem:

1. der durch die Ausdehnung der Eisbedeckung vor allem auf der Nordhalbkugel vergrösserte Temperaturgradient zwischen den Tropen und den gemässigten Breiten,
2. die erwähnte Verschiebung der äquatorialen Trockengrenzen und
3. die in der Gegenwart beobachtete Erscheinung, dass sich Änderungen in der aussertropischen Zirkulation auf den Passatkreislauf und umgekehrt übertragen.

Zusammenfassend kann somit festgestellt werden, dass die allgemeine Zirkulation der Atmosphäre während einer Glazialepoche infolge der grösseren Eisbedeckung in allen Teilen verstärkt und der aussertropische Kreislauf besonders auf der Nordhalbkugel äquatorwärts ausgedehnt waren. Es ist sehr wahrscheinlich, dass damals wegen der starken nordhemisphärischen Vereisung nicht wie heute die südliche Zirkulation über den Äquator nach Norden übergriff, sondern eher umgekehrte Verhältnisse geherrscht haben.

2. Atmosphärische Zirkulation und Eiszeitursachen

Die auf Grund zahlreicher Zeugnisse der Diluvialgeologie und des diluvialen Klimas abgeleitete Verstärkung der allgemeinen Zirkulation während der Glazialperioden gegenüber der Jetztzeit steht, worauf bereits des öfteren hingewiesen worden ist, im Widerspruch zu den in der jüngsten Vergangenheit beobachteten Zusammenhängen zwischen Zirkulations-, Klima- und Gletscherschwankungen.

Nach den Untersuchungen von WAGNER (1929) und SCHERHAG (1936, 1939) hat bekanntlich seit etwa 1850 eine zunehmende Verstärkung der atmosphärischen Zirkulation stattgefunden. Von den damit verbundenen Klimaänderungen und sonstigen Erscheinungen sind hier besonders zu erwähnen:

1. Erniedrigung der Sommer- und Erhöhung der Wintertemperaturen, d. h. Abschwächung der jährlichen Temperaturschwankungen in Mittel- und Nordeuropa;
2. Vermehrung der Niederschläge im gleichen Raum mit Ausnahme eines Gebietes zwischen Grönland-Mittelengland-Bottnischer Meerbusen-Nordkap und
3. Verminderung der Niederschläge in den Rossbreiten;
4. Nordwärts-Verlagerung des Azoren-Maximums,
5. zunehmende Austrocknungserscheinungen am polaren Rand des nördlichen Trockengürtels,
6. Rückgang der Gletscher fast aller Hochgebirge,
7. starker Rückgang der arktischen Eisbedeckung.

Demgegenüber waren in den Vereisungszeiten bei verstärkter Zirkulation die Temperaturen erniedrigt, das Azorenhoch eher südwärts als nach Norden verlagert, die polaren Ränder der Wüstengürtel feuchter als heute, und vor allem war die Schneegrenze um 500—1 500 m gegenüber der Jetztzeit gesenkt, die Landgebiete der mittleren und hohen Breiten waren vereist und sämtliche Hochgebirge weitaus stärker vergletschert als heute.

Zu diesen Widersprüchen muss festgestellt werden, dass ein Vergleich zwischen der oben abgeleiteten glazialen und der gegenwärtigen Zirkulationsverstärkung aus kausalen Gründen überhaupt nicht möglich ist. Die eiszeitliche Intensivierung des Luftkreislaufes wurde festgestellt auf Grund der »beobachteten« Verhältnisse während eines Eishöchststandes. Sie ist erst eine Folge der zunehmenden Ausdehnung der Eisgebiete in den mittleren Breiten, d. h. eine durch terrestrische Einflüsse sekundäre Veränderung der ursprünglichen Zirkulation. Dagegen wurde die Zirkulationsschwankung der Gegenwart sehr wahrscheinlich durch extraterrestrische Vorgänge (n. WAGNER durch

eine Vermehrung der solaren Einstrahlung) ausgelöst.

Mit der Frage, welche Zirkulationsverhältnisse in der primären Phase einer Glazialperiode herrschten, kommt man notwendigerweise zum Problem der Vereisungsursachen. Die erwähnten Beobachtungen der Gegenwart zeigen eindeutig, dass eine gegen heute verstärkte Zirkulation weder als direkte noch als mittelbare Ursache der Vereisungen betrachtet werden kann. Man muss im Gegenteil feststellen, dass derartige Verhältnisse durch den vermehrten Wärmetransport in die höheren Breiten vereisungsfeindlich sind.

Leider ist es bisher nicht gelungen, die ursächlichen Zusammenhänge zwischen Gletscher- und Klimaschwankungen einwandfrei zu ermitteln. Die von KÖPPEN-WEGENER (1924) in Verbindung mit der astronomischen Theorie der Eiszeiten eingeführte Vorstellung, dass ein ozeanisches Klima mit erhöhten Niederschlägen und verringerter jährlicher Temperaturamplitude eine Zunahme der Vergletscherung begünstige, erscheint zwar durchaus einleuchtend, steht jedoch im Widerspruch mit den Erfahrungen der jüngsten Vergangenheit. WAGNER nimmt daher an, dass die Gletscherbewegungen in erster Linie an die mittleren Jahrestemperaturen gebunden sind, die offenbar auf der ganzen Erde seit etwa 1800 zunehmen.

Wenn somit die gegenwärtigen Beobachtungen zeigen, dass eine Verstärkung des Luftkreislaufes mit Gletscherrückgang verbunden ist, so muss daraus gefolgert werden, dass in den jeweiligen Zeiten des Beginns eines Eisvorstosses die Zirkulation relativ zur Gegenwart abgeschwächt, d. h. der Luftaustausch zwischen den hohen und subtropischen Breiten vermindert war. Schwache Zirkulation bedeutet aber nach den Untersuchungen von DEFANT, SCHERHAG, SCHEDLER u. a.: Häufung kalter Winter in hohen und mittleren Breiten und Verlagerung vor allem der winterlichen Zyklonentätigkeit auf die südlicheren Bahnen. Damit wird die Frage nach der zur Vereisung der grossen Landgebiete in den hohen und mittleren Breiten erforderlichen kräftigen Feuchtezufuhr problematisch. Eine Klärung dieser Widersprüche ist nur unter Berücksichtigung der geographischen Verhältnisse der vergletschernden Gebiete möglich. Es soll nun versucht werden, die Mitwirkung und Bedeutung der allgemeinen Zirkulation bei der Entstehung der diluvialen Eisgebiete kurz zu skizzieren, wobei als Ausgangszustand (z. B. Riss-Würm-Interglazial) bezüglich der Land-Meer-Verteilung, Orographie und polaren Eisbedeckung etwa die heutigen Verhältnisse angenommen werden.

a. *Europäischer Eisschild.* Bei der Rekonstruktion der Vergletscherung Nordeuropas treten die soeben erwähnten Fragen besonders deutlich in Erscheinung. Nach den in der Gegenwart beobachteten Zusammenhängen zwischen Gletscherausdehnung und Zirkulation kann die Bildung der skandinavischen Vereisung nur bei einem im Vergleich zur Jetztzeit geschwächten Luftkreislauf begonnen haben:

Die geringe Wärmeadvektion in die hohen Breiten musste dabei zunächst zu einer Erniedrigung der Temperaturen im Nordpolarbecken und schliesslich zu einer Ausdehnung der arktischen Meereisbedeckung möglicherweise bis nahe an die nordeuropäische Küste führen. Im nordskandinavischen Hochgebirge wurde zunächst die infolge schwacher Zirkulation bewirkte Abkühlung durch die Kältewirkung der vergrösserten Polareiskappe verstärkt und damit eine Senkung der Schneegrenze bewirkt. Gleichzeitig erfolgte aber auch eine Verschärfung der meridionalen Temperaturgegensätze im nordatlantischen Raum. Der Gleichgewichtszustand zwischen Temperaturverteilung und Zirkulation wurde gestört, und unter langsamen Schwankungen — ähnlich den von DEFANT (1924) untersuchten Zirkulationsschwankungen infolge vulkanischer Lockerausbrüche — fand eine Intensivierung des Luftaustausches statt. Damit wurde zwar die Zyklonentätigkeit in den nördlichen Breiten verstärkt, durch vermehrte Wärmeadvektion aber die Schneegrenze wieder gehoben und die Packeisgrenze zurückgedrängt. Diese Schwierigkeit verschwindet jedoch, wenn man bedenkt, dass die diluviale Klimaverschlechterung nach allen vorliegenden Zeugnissen eine weltweite Erscheinung war, die (ohne Eiswirkung) eine generelle primäre Temperaturabnahme auf der Erde von mindesten 4°—7° C zur Folge hatte. Dieser Temperaturrückgang unterstützte den gesamten dargestellten Vorgang, so dass die oben erwähnte Zirkula-

tionsverstärkung in ihrer erwärmenden Wirkung für die hohen Breiten sicher nur wenig in Erscheinung trat, aber doch die Feuchtezufuhr nach Nordeuropa vermehrte und ein Anwachsen der skandinavischen Gletscher ermöglichte.

Der wesentlichste Punkt bei der Entwicklung einer stärkeren Vergletscherung des skandinavischen Hochgebirges ist somit das Zusammentreffen von drei Abkühlungsursachen: 1. genereller Temperaturrückgang auf der Erde, 2. Temperaturerniedrigung durch verminderte Zirkulation im Nordpolarbecken, 3. Kältewirkung der vergrösserten polaren Eiskappe.

Für die weitere Entwicklung der Gletscher- und Eisausbreitung wird der zuerst von Brooks (1925) herangezogene und später von Klute, Behrmann und Wundt hervorgehobene Vorgang der Selbstverstärkung einer Eiskappe der ausschlaggebende Faktor gewesen sein. Je weiter südwärts die Eismassen vordrangen, desto kräftiger wurde die Zirkulation, bis die Wärmezufuhr aus den niederen Breiten gemeinsam mit den strahlungsbedingten Temperaturverhältnissen im Eisrandgebiet die Kältewirkung des Eises kompensierten und damit eine Stillstands- bezw. Endlage der Gletscherausdehnung erreicht war. Man darf jedoch weder die Ausbreitung des Eises als einen absolut stetigen Vorgang noch die Endlage als einen absoluten Ruhezustand ansehen. Der Eisrand führte vielmehr ständige Pendelbewegungen aus, deren Ursachen (mit entsprechender Verzögerung) wahrscheinlich in langperiodischen Zirkulationsschwankungen ähnlich den in historischer Zeit stattgefundenen zu suchen sind.

b. *Nordamerikanische Vereisung.* In Nordamerika bestanden während der Diluvialzeit drei Eiszentren — Kordillerenvereisung, Keewatin- und Labradoreisschild —, deren maximale Ausdehnung nicht gleichzeitig, sondern nacheinander, offenbar von West nach Ost fortschreitend, erfolgte (Coleman, Leverett, Antevs). Die Lage der beiden letztgenannten Zentren im Innern bezw. an der Ostküste der nordamerikanischen Landmasse lässt erkennen, dass das für die skandinavische Vereisung entworfene Bildungsschema hier nicht einfach übertragen werden kann. Nur die Kordillerenvergletscherung als ältester Eisschild weist eine ähnliche geographische Lage auf wie der nordeuropäische, allerdings fehlt auch hier die direkte Berührung mit dem arktischen Becken. Sie findet die einfachste Erklärung:

Der allmähliche Temperaturrückgang im Pliozän bewirkte ein Absinken der Schneegrenze, wobei die Gletscherausdehnung noch begünstigt wurde durch die damals wahrscheinlich gegenüber der Jetztzeit weiter nordwärts gelegene Hauptzugstrasse der Depressionen (Brooks 1924).

Nachdem die Kordillerenvereisung ihr Maximum bereits überschritten hatte, entwickelte sich der Keewatineisschild in dem westlich der Hudsonbay gelegenen Tiefland (300—500 m NN), welches sich heute durch ein exzessives Kontinentalklima mit nur 300—350 mm Jahresniederschlag auszeichnet. Zur Bildung dieses Eisgebietes musste die Schneegrenze erheblich abgesunken sein. Es war also eine starke Temperaturerniedrigung (Kältewirkung des angrenzenden Kordillereneises) und eine gegen heute vermehrte Feuchtezufuhr, d. h. eine Verstärkung der Zyklonentätigkeit auf südlicheren Bahnen, erforderlich.

Wahrscheinlich wird diese Zirkulationsverstärkung bereits ein sekundärer Vorgang, eine Folge der zunehmenden Eisbedeckung des arktischen Archipels und der damit verbundenen Südwärtsverlagerung der Zyklonenbahnen gewesen sein. Dabei musste, in Übereinstimmung mit den Beobachtungen, die Kordillerenvereisung im nördlichen Teil wegen der abnehmenden Feuchtezufuhr und im küstennahen Raum infolge des vermehrten Wärmetransportes allmählich zurückgehen. Die niederschlagsbringenden Winde werden zunächst vorwiegend aus westlicher Richtung (Enquist, Antevs), später jedoch auch von der atlantischen Seite gekommen sein.

Das Labradorzentrum bildete sich wenig nach dem Keewatineis. Seine Mächtigkeit betrug im Mittel nur 300—500 m, und grosse Teile Neufundlands sowie des Torngate-Tafellandes blieben völlig eisfrei. Das deutet offenbar auf eine gegenüber den anderen Zentren schlechtere Ernährung dieses Eisgebietes. Trotzdem muss die Feuchtezufuhr damals grösser gewesen sein als heute (400—800 mm Jahresniederschlag). Wie bereits Enquist hervorgehoben hat, müssen die Schneewinde aus dem östlichen Quadranten, d. h. von der

Rückseite der südwärts verlagerten nordatlantischen Depression, vom Nordatlantik gekommen sein.

Die Labradorvereisung stellt also eine sekundäre Bildung bei stark erniedrigten Temperaturen (allgemeiner Temperaturrückgang und Kältewirkung des Keewatineises) nach erfolgter Zirkulationsverstärkung dar. Letzteres stimmt gut mit den Beobachtungen aus der Gegenwart überein, wonach in Labrador im Laufe der letzten 30—40 Jahre die jährliche Niederschlagsmenge und die Wintertemperaturen zugenommen, die Jahrestemperaturen jedoch abgenommen haben.

Überblickt man die nordamerikanische Vereisung in ihrer Gesamtheit, so ergibt sich der gleiche Entstehungsvorgang wie bei der Vergletscherung Skandinaviens: 1. Auslösung der Vereisung als Folge eines allgemeinen Temperaturrückganges und verminderter Zirkulation; 2. Weitere Ausbreitung des Inlandeises, d. h. Bildung des Keewatin- und Labradorzentrums, durch zusätzliche Abkühlung (Kältewirkung des Eises) bei Intensivierung des allgemeinen Luftkreislaufes.

c. *Nordasiatische Vereisung.* Über die zur Bildung der nordasiatischen Eisgebiete erforderlichen atmosphärischen Verhältnisse sind nur allgemeine Aussagen möglich, weil die zeitliche Stellung der einzelnen Vereisungszentren und ihre Ausdehnung noch nicht genügend geklärt sind. Da es sich hier grösstenteils um Gebiete eines gegenwärtig polaren innerkontinentalen Trockenklimas handelt, muss neben dem primären Temperaturrückgang — der zur Vereisung des Nordpolarbeckens bis an die nordsibirische Küste führte- und der Kältewirkung dieser polaren Eiskappe die Feuchtezufuhr verstärkt gewesen sein. Dabei konnten die schneebringenden Luftströmungen nur aus west- bis südwestlicher Richtung, d. h. aus dem nordatlantischen Verdunstungsgebiet kommen. Die Bildung der nordasiatischen Eisgebiete konnte demnach erst in der sekundären Phase der nordeuropäischen Vergletscherung bei verstärkter Zirkulation (weites Vordringen der über das Mittelmeer-Schwarzmeer nach Nordosten ziehenden Depressionen) erfolgen.

d. *Antarktika.* Die Vereisung des Südpolarkontinentes ist wohl das schwierigste Problem der gesamten Glazialforschung, da die rezente Eiskappe nahezu sämtliche vorpleistozänen Ablagerungen bedeckt und keine Aussagen über den Zeitpunkt des Vereisungsbeginnes zulässt. Die im antarktischen Randgebiet (Umgebung vom Grahamland) gefundenen tertiären Floren lassen ähnlich wie auf der Nordhemisphäre einen zunehmenden Temperaturrückgang im Laufe des Neogen erkennen. Die pliozäne Steppen- und Tundrenflora der Cockburn-Insel lässt sogar eine mindestens den zentralen Teil von Antarktika bedeckende Vereisung vermuten. Aussagen über die zur Bildung des Inlandeises erforderlichen meteorologischen Verhältnisse sind jedoch nicht möglich, zumal auch die Landverteilung im Bereich des spättertiären Südpolarkontinentes unbekannt ist.[1]

Demgegenüber konnte die an mehreren Stellen von Antarktika beobachtete grössere Mächtigkeit der Eisbedeckung in jüngerer geologischer Vergangenheit durch MEINARDUS (1928) befriedigend erklärt werden: Ein Eiszuwachs im Gebiet des südpolaren Inlandeises ist nur durch eine vermehrte Feuchtezufuhr möglich. Eine Erhöhung des Wasserdampfgehaltes der Luft konnte aber in den Glazialzeiten nur durch vermehrte Verdunstung infolge Zunahme der mittleren Windgeschwindigkeit in der südlichen Westwinddrift, d. h. durch Intensivierung der allgemeinen Zirkulation, erfolgen. — Auch hier ist die zeitliche Einordnung des Eishöchststandes noch problematisch, so dass eine Parallelisierung mit den nordhemisphärischen Verhältnissen nur eine reine Vermutung darstellen würde.

Zusammenfassend kann festgestellt werden, dass — soweit zuverlässige Aussagen möglich sind (vor allem im nordeuropäischen Eisgebiet und bei der nordamerikanischen Vereisung) — die atmosphärischen Verhältnisse in der primären Phase einer Glazialzeit bezw. des Diluviums ausgezeichnet waren durch

1. einen allgemeinen Temperaturrückgang auf der Erde,
2. gegenüber der Jetztzeit verminderte Zirkulation und
3. eine zusätzliche Abkühlung durch die Eisbedeckung des Nordpolarbeckens (gilt nur für die Nordhemisphäre).

[1] Die Diskussion der vorhandenen Möglichkeiten muss einer späteren Arbeit vorbehalten bleiben.

3. Eiszeittheorien und Zirkulation

Der nächste Schritt ist eine Untersuchung der zur Erklärung der diluvialen Klimaschwankungen aufgestellten Theorien hinsichtlich des mindestens für die Nordhalbkugel gültigen Kriteriums, dass zur Auslösung eines Eisvorstosses eine allgemeine Temperaturabnahme und eine gegen heute geschwächte Zirkulation erforderlich ist. Da es müssig wäre, sämtliche, bisher aufgestellte Eiszeittheorien zu besprechen, sollen nur die wichtigsten Punkte der 4 gegenwärtig zur Diskussion stehenden Gruppen (solare — kosmische — geographische — astronomische Theorie), deren Grundlagen vorausgesetzt werden müssen, behandelt werden.

A. *Solare Ursachen.* Der nächstliegendste und daher wohl älteste Versuch zur Deutung des Eiszeitphänomens ist die Annahme einer Verminderung der solaren Strahlungsenergie, wodurch zunächst der allgemeine Temperaturrückgang auf der Erde erklärt wird. Damit muss aber gleichzeitigauch die Intensität der atmosphärischen Zirkulation geschwächt worden sein, denn die Berechnungen der Veränderung des zonalen Wärmegefälles bei einer Änderung der Solarkonstanten durch ZENKER, LIZNAR und HOPFNER (Zusammenfassung bei KERNER (1940) zeigen trotz zahlenmässiger Unterschiede prinzipiell das gleiche Ergebnis: Die äquatorialen Gebiete werden von der Strahlungsänderung stets stärker betroffen als die höheren Breiten. Bei einer Verminderung der Zustrahlung werden also die meridionalen Bestrahlungsdifferenzen geschwächt und damit die Zirkulation vermindert. Der Vorgang wird zwar durch sekundäre Veränderungen noch etwas kompliziert (Verringerung der Verdunstung und Bewölkungsrückgang in den Tropen und gemässigten Breiten; in den Subtropen durch Schwächung der Absinkbewegungen Bewölkungszunahme) es muss sich aber schliesslich ein Gleichgewicht zwischen Einstrahlung und Temperaturverteilung einstellen, bei dem gegenüber der vorangegangenen Zeit die Temperaturen erniedrigt sind und der meridionale Luftaustausch geschwächt ist.

Zum gleichen Ergebnis führt die von DEFANT (1921) aufgestellte Differentialgleichung über den Zusammenhang von Einstrahlung (E), Ausstrahlung (A), Luftzirkulation (Z) und Temperatur (T):

1. E + = Z +, T + 3. A + = Z +, T—
2. E — = Z —, T — 4. A — = Z —, T +

In der primären Vereisungphase wäre danach der Vorgang 2 anzusetzen, in der sekundären — durch die von WUNDT (1938a) infolge verstärkter Strahlungsreflexion des Eises abgeleitete Erhöhung der Erdalbedo um etwa 4 % — dagegen Vorgang 3.

Die Annahme einer Verminderung der solaren Zustrahlung, deren Gründe hier nicht zur Diskussion stehen, führt also zu den zur Einleitung einer Vereisungsphase nötigen Bedingungen. Demgegenüber hat C. G. SIMPSON (1934) versucht, nachzuweisen, dass auch bei Verstärkung der Sonnenstrahlung ein Eisvorstoss in den höheren Breiten möglich sei.

Nach dieser Theorie fand im Diluvium eine zweimalige Verstärkung und Abschwächung der solaren Strahlungsintensität statt, wobei in den Zeiten zunehmender und abnehmender Strahlung Eisvorstösse — entsprechend den 4 Glazialperioden — ausgelöst wurden, während die Zeiten maximaler und minimaler Strahlung eisfrei, d. h. Interglaziale, waren. Daraus folgt, dass sich die Glaziale 1 und 3 in Zeiten zunehmender, die Glaziale 2 und 4 dagegen bei abnehmender Intensität der Zirkulation entwickelten. Von den Interglazialen waren das erste und dritte (Strahlungsmaxima) warm-feucht mit sehr kräftiger Zirkulation, das zweite und die Gegenwart dagegen kalttrocken mit schwachem Luftaustausch.

Zu den zahlreichen von geologischer und meteorologischer Seite vorgebrachten Einwänden gegen die Simpson'sche Theorie (WAGNER, KÖPPEN, BROOKS) sei im Rahmen des vorliegenden Themas noch hinzugefügt:

1. Die Entstehung der Eisvorstösse 1 und 3 bei (primär) verstärkter Zirkulation steht in krassem Widerspruch zu den in der Gegenwart gemachten Beobachtungen.
2. Die nordafrikanischen Pluviale werden in die warm-feuchten Interglaziale verlegt. Da in diesen Zeiten die Zirkulation kräftig entwickelt gewesen sein muss, ist nicht einzusehen, wie die verstärkte Niederschlagstätigkeit im Mittelmeergebiet zustande kommen konnte. (Azorenmaximum nordwärts verlagert, Depressionen bevorzugen nördliche Zugstrassen!)

Zusammenfassend muss man feststellen, dass der Theorie von SIMPSON wenig Wahrscheinlichkeit zukommt, obwohl manche Einzelheiten, z. B. die sekundären Wirkungen der veränderten hydrometeorischen Verhältnisse, wertvolle neue Gesichtspunkte im allgemeinen Eiszeitproblem eröffneten.

B. *Kosmische Ursachen.* Von den Theorien, die kosmische Ursachen zur Erklärung des eiszeitlichen Temperaturrückganges heranziehen, ist nur die Anschauung von HIMPEL (1947) bedeutsam. Ihre Erörterung erübrigt sich jedoch, da sie im wesentlichen nur die im Diluvium mehrfach erfolgten Schwächungen der solaren Strahlungsenergie zu erklären sucht und sich daher bezüglich der irdischen Temperatur- und Zirkulationsverhältnisse keine neuen Gesichtspunkte ergeben.

C. *Geographische Ursachen.* In der Gruppe der geographischen Theorien wird versucht, die diluvialen Klimaänderungen allein durch Veränderung der geographischen Verhältnisse (Land-Meer-Verteilung, Orographie usw.) ohne Strahlungsänderungen zu erklären. Es gilt somit zu untersuchen, ob die unmittelbar prädiluvialen Veränderungen des Reliefs der Erdoberfläche ausreichten, um über die dadurch erzwungenen Änderungen der Zirkulation und des Klimas die Vereisungen auszulösen.

Hierbei sind die Untersuchungen von BROOKS (1925) und KERNER (1932) über die Frage des akryogenen Polarklimas von wesentlicher Bedeutung. Unter gegenwärtigen Strahlungsverhältnissen ist ein eisfreies Nordpolarmeer nur bei kräftigem meridionalen Wärmeaustausch (Vorhandensein mehrerer breiter Meeresverbindungen mit den tropischen Warmwassergebieten) und starker Wellenbewegung der Wasseroberfläche im Polarbecken möglich. Die allgemeine Zirkulation muss also kräftig entwickelt sein. Sobald dieser Luftaustausch infolge von Veränderungen der Land-Meer-Verteilung (z. B. Abschliessung des Nordpolarbeckens durch ausgedehnte zirkumpolare Landgebiete) vermindert oder behindert wird, sinkt die Temperatur im Polarbecken unter den für das akryogene Seeklima kritischen Wert, es bildet sich eine polare Meereiskappe, die durch ihre Kältewirkung zur weiteren Abkühlung der umgebenden Landgebiete beiträgt (Selbstverstärkung usw).

Die zahlreichen Funde wärmeliebender, z. T. subtropischer Pflanzen aus dem Alttertiär Grönlands und Spitzbergens machen mindestens für das Eozän das Vorhandensein eines weitgehend eisfreien Nordpolarbeckens wahrscheinlich. Vergleicht man die paläogeographische Karte des Eozän von MATTHEW (3 breite meridionale Meeresverbindungen im Gebiet des Nordatlantik, der Beringstrasse und zwischen Ural und Ob) mit der Karte des Pliozän von KOKEN-ARLDT (Abschnürung des Nordpolarmeeres), so treten die von BROOKS und KERNER geforderten Verhältnisse deutlich in Erscheinung.

Obwohl derartige Vergleiche bei der Unsicherheit paläogeographischer Rekonstruktionen nur mit Vorbehalt angewendet werden dürfen, kann man es doch als gesichert ansehen, dass zu Beginn des Diluviums die Landverteilung in den hohen Breiten etwa der heutigen entsprach. Zudem weist die Floren- und Faunenentwicklung im Neogen eindeutig auf eine zunehmende Abkühlung in den mittleren und hohen Nordbreiten.

Die angegebenen Veränderungen der nordhemisphärischen Festlandskonfiguration könnten somit die für den Beginn der Vereisung erforderlichen meteorologischen Voraussetzungen (Abschwächung der allgemeinen Zirkulation) geschaffen haben, wobei sie durch die im Rahmen der tektonischen Phasen erfolgten Gebirgsbildungen und Hebungen unterstützt wurden. — Das Kernproblem bilden dabei die Verhältnisse im nordatlantischen Raum, wo die infolge der verstärkten Vereisung des Nordpolarmeeres sekundär verstärkte Zirkulation der weiteren Eisausdehnung entgegenwirkte. Es fehlt hier die oben erwähnte dritte Komponente der Temperaturerniedrigung. Nimmt man mit BUBNOFF, JESSEN, KERNER, QUIRING u. a. die Existenz der durch prädiluviale Hebung entstandenen Island—Färöer—Schwelle an, so ist das Problem zunächst gelöst, da auf diese Weise der Skandik und das Nordpolarbecken von der warmen nordatlantischen Drift abgesperrt wurden. Zur Erklärung der vier diluvialen Vereisungsperioden muss dann aber ein viermaliges Heben und Senken dieser Schwelle angenommen werden, was zweifellos recht unwahrscheinlich ist (BEHRMANN 1948).

Einen Versuch des Ausgleiches unternimmt WUNDT (1944), indem er nur eine Behinderung (nicht völlig Absperrung) des Golfstromes durch eine zwar gegen heute höhere, aber noch untermeerische Islandbarre annimmt.

Man muss somit zu dem Ergebnis kommen, dass die stattgefundenen geographischen Veränderungen zweifellos massgeblich am Zustandekommen der für die Vereisungen erforderlichen meteorologischen Verhältnisse beigetragen haben, dass sie aber nicht als alleinige Ursache der Glazialperioden angesehen werden können.

D. *Astronomische Theorie.* Bei der astronomischen Theorie der Eiszeiten werden die klimatischen Wirkungen der durch die Schwankungen der Erdbahnelemente ausgelösten Änderungen der solaren Bestrahlungsverhältnisse als Hauptfaktoren für die Entstehung der Vereisungen herangezogen.[1] Die sich hier ergebenden meteorologischen Verhältnisse sind wohl am besten von WUNDT (1944) dargestellt worden.

Die Grundlage für die Parallelisierung von Strahlungsschwankungen und Glazialperioden bildet die Annahme von KÖPPEN und WEGENER (1924), dass milde Winter (kräftige Zyklonentätigkeit, starke Schneefälle in mittleren und hohen Breiten) und vor allem kühle Sommer (geringe Ablation), d. h. ein allgemein ozeanisches Klima vereisungsgünstig sind. Dementsprechend wurden die Glazialperioden mit den Zeitabschnitten geringer sommerlicher Bestrahlung bezw. geringer jahreszeitlicher Bestrahlungsdifferenzen identifiziert (Steillage der Erdachse, Perihellage im Winter bei starker Excentrizität der Erdbahn). Durch die gleiche Kombination der Bahnelemente, welche die jahreszeitlichen Unterschiede schwächt, werden aber die meridionalen Differenzen verstärkt, woraus folgt, dass — sofern die Änderung der meridionalen Bestrahlungsunterschiede eine gleichsinnige Veränderung der allgemeinen Zirkulation zur Folge hat — die primären Phasen der Glazialperioden an Zeiten verstärkter Zirkulation gebunden waren.

[1] Es wird im folgenden nur die von MILANKOVITCH entwickelte Theorie behandelt. Die von SPITALER ausgearbeitete Fassung muss wegen der klimatologisch unhaltbaren Definition der mittleren Bestrahlung eines Breitenkreises abgelehnt werden.

Zu diesen hier kurz skizzierten Zusammenhängen ist folgendes festzustellen:

1. Wie bereits erwähnt, ist es noch unentschieden, ob kalte Sommer und eine Verminderung der jährlichen Temperaturschwankung die Gletscherausbreitung begünstigen.
2. Kühle Sommer sind in Nordeuropa nach den Erfahrungen an schwache Zirkulation, d. h. geringe Zyklonentätigkeit, geringe Ausdehnung des Azorenhochs und Verarmung der nördlichen Zugstrassen, gebunden.
3. Die gegenwärtigen Zirkulationsschwankungen und ihre Folgen bezüglich der Gletscherbewegung und Vereisung des Nordpolarmeeres stehen im offenen Widerspruch zu der Zirkulationsverstärkung nach der astronomischen Theorie.

Zur genaueren Untersuchung des letztgenannten Punktes wurden nach den Strahlungstabellen von MILANKOVITCH (1930) die Bestrahlungsdifferenzen zwischen 35° und 65° sowie zwischen 25° und 75° Breite jeweils für die als Hauptglazial- und Interglazialperioden bezeichneten Jahre bestimmt und mit den entsprechenden Werten der Gegenwart verglichen. Unter der bereits erwähnten Voraussetzung des Zusammenhanges zwischen Strahlungsgefälle und Luftkreislauf ergaben sich relativ zur Gegenwart folgende Zirkulationsänderungen:

1. Glazialzeiten der Nordhalbkugel:

Zirkulation im	Nordsommer	+
	Nordwinter	+
Zirkulation im	Südsommer	+
	Südwinter	—

2. Interglazialzeiten der Nordhalbkugel:

Zirkulation im	Nordsommer	+
	Nordwinter	—
Zirkulation im	Südsommer	—
	Südwinter	+

Die Zusammenstellung zeigt, dass nach der astronomischen Theorie während der nordhemisphärischen Glaziale sowohl die Sommer- als auch die Winterzirkulation (primäre Änderungen!) verstärkt waren, was nicht als vereisungsgünstig angesehen werden kann. Demgegenüber sind die Zirkulationsverhältnisse in den Interglazialen eher zutreffend, da für die hohen Breiten schwacher Austausch im Winter geringere Schneezufuhr und kräf-

tiger sommerlicher Austausch starke Ablation zur Folge haben.

Die aufgezeigten Schwierigkeiten hinsichtlich der primären Intensivierung des atmosphärischen Kreislaufs in den Glazialperioden hat WUNDT (1944) durch Kombination der astronomischen Theorie mit Reliefänderung zu lösen versucht: Die Hebung der Island—Färöer—Brücke gegen Ende des Tertiär versetzte die hohen Nordbreiten in einen Zustand der Vereisungsbereitschaft (primäre Abkühlung), während der zur Auslösung der Vereisungen erforderliche kritische Zustand durch die Bestrahlungsänderungen infolge Schwankungen der Erdbahnelemente (sommerliches Strahlungsminimum als zweite temperaturerniedrigende Komponente) erreicht bezw. überschritten wurde, wozu sich als dritter Faktor noch der Vorgang der Selbstverstärkung gesellte. Diese Verknüpfung zweier Ursachen kann als sehr brauchbar angesehen werden, denn bei entsprechend veränderten geographischen Verhältnissen können primäre Abkühlung und verstärkte, aber z. T. behinderte Zirkulation zweifellos vereisungsfördernd gewirkt haben. Andererseits wird die bei alleiniger Anwendung der Reliefveränderung als Eiszeitursache erforderliche mehrmalige Hebung und Senkung der Schwelle umgangen, da, sobald die geographischen Voraussetzungen vorhanden sind, die innere Struktur des Eiszeitalters durch die Schwankungen der Bahnelemente bestimmt wird.

Eine entsprechende Untersuchung der jahreszeitlichen Zirkulationsveränderungen auf der Südhalbkugel[1] ergab, dass zur Zeit der nordhemisphärischen Strahlungsminima (= Glaziale) im Süden der sommerliche Kreislauf verstärkt und der winterliche allgemein abgeschwächt war, während zur Zeit der südlichen Strahlungsminima umgekehrte Verhältnisse geherrscht haben, wobei im Mittel die winterliche Intensivierung stärker war als die sommerliche Schwächung. Eine Deutung dieser Verhältnisse ist schwierig, zumal, wie erwähnt, Angaben über den Zeitpunkt des Beginns der antarktischen Vereisung und die dabei vorhandenen geographischen Bedingungen fehlen. Sofern der Südpolarkontinent bereits vereist war, konnte nach MEINARDUS (1937) nur eine Verstärkung der Zirkulation eine Zunahme der Eisbedeckung bewirken. Dabei wird jedoch m. E. eine sommerliche Verstärkung (Wasserdampfzufuhr) wirksamer gewesen sein als eine winterliche. Andererseits wird für die Bildung des in der Zone der Westdrift gelegenen patagonischen Inlandeises eine winterliche Intensivierung günstiger gewesen sein (verstärkter Schneefall).

Überblickt man abschliessend die vorangegangenen Erörterungen, so ergibt sich, dass ausser den solaren Theorien keine andere Hypothese allein in der Lage ist, die eiszeitlichen Verhältnisse zu erklären. Andererseits steht bei den ersteren noch immer die Frage nach der Ursache der wiederholten Verringerung der Solarkonstanten offen. Eine Ausnahme bildet hier die kosmische Theorie von HIMPEL, die recht gut fundiert erscheint, jedoch auch noch einiger astronomischer Untermauerung bedarf. Ganz allgemein muss festgestellt werden, dass es für den gegenwärtigen Stand der Eiszeitforschung wichtig ist, die in den verschiedenen Theorien dargelegten Gedanken zu kombinieren. Die von WUNDT durchgeführte Vereinigung von Reliefänderungen, Schwankungen der Erdbahnelemente und Selbstverstärkung war zweifellos recht fruchtbar. Nicht das dogmatische Festhalten an einer einmal aufgestellten Hypothese, sondern die Verbindung der Theorien untereinander zur möglichst vollständigen Erklärung aller eiszeitlichen Erscheinungen und ihre Anpassung an die sich ständig vermehrenden Erkenntnisse müssen das Ziel der heutigen Diluvialforschung sein.

[1] Die durch die Bahnelemente bewirkten Strahlungsschwankungen sind bekanntlich auf den beiden Hemisphären nicht genau synchron. Die im Mittel 8—11 Jahrtausende betragenden Zeitdifferenzen zwischen dem Eintritt der Strahlungsminima auf Nord- und Südhemisphäre dürften jedoch von MEINARDUS (Geol. Rdsch. 34 — 1944 — Klimaheft) befriedigend erklärt worden sein.

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The Behavior of Jet Streams over Eastern North America during January and February 1948

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Abstract

A graph of the latitude of jet streams at 80° W longitude for a 49-day period in the winter of 1947—8 shows that the jet stream undergoes changes which appear as discontinuities on such a graph. Several cases are discussed which illustrate the manner in which these discontinuities appear. The new jets so formed then appear to experience a southward displacement before their disappearance. Accompanying the southward displacement at this longitude is a disappearance of the solenoidal field accompanying the jet in the lower troposphere, and an increase of the potential temperature in the center of the jet.

Introduction

Most previous detailed studies of the so-called "jet stream" (Palmén and Newton 1948, Riehl 1948, Hovmöller 1948) have been concerned primarily with the structure of the jet and its relation to other atmospheric phenomena *in selected situations*. However, even in some of these studies there are indications that the jet is not a simple phenomenon. Palmén and Newton (1948) averaged conditions at 80° W for selected cases, using the polar front as a moving reference point to obtain a meridional cross section. There is definite evidence on their mean cross section of a secondary wind maximum south of the primary maximum associated with the front — this in spite of their method of averaging, which would tend to suppress any but the primary maximum. There is also some evidence in Riehl's study (1948) (e. g. fig. 2 b therein) of a double maximum of wind.

Some phases of this more complicated nature of the jet stream have been discussed recently by Cressman (1950). He found that (at least during certain periods) there seemed to be a tendency for a hemispheric jet stream to move from high to low latitudes with a new current forming in the North as the previous maximum moved southward. He also found that there was a tendency for the individual jets to attain their maximum strength as they reached middle latitudes and to increase their elevation and potential temperature as they moved southward. The rate of southward progression was of the order of one-half degree of latitude per day.

In connection with an attempt to compute the transfer of energy from west to east across the longitude of 80° W, Dr. H. L. Kuo and the writer analysed twice-daily meridional cross sections for that longitude for the 49-day

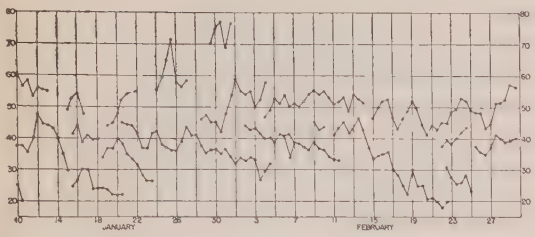


Fig. 1. Graph of the latidude of jets at longitude 80° W as a function of time for the period 10 Jan.—29 Febr. 1948.

period 10 January — 29 February 1948. These cross sections included an analysis of the wind field normal to the cross section, using both observed and gradient winds (geostrophic winds corrected for path curvature). Although the unavoidable errors in the energy transfer calculations rendered those computations meaningless, it was soon found that a southward shift of the wind maxima seemed to be indicated by the successive cross sections. Because of the interesting results uncovered by CRESSMAN at that time, it was decided to re-examine this material with CRESSMAN's results in mind. The continuity of the various maxima from day to day was then put on a more definite basis by further analysis and examination of the twice-daily 500- and 300-mb charts for the region of North America during the entire period. In some cases additional cross sections were analysed in other regions of the continent.

The latitudinal movement of the wind maxima at 80° W during the period is summarized in fig. 1. Each dot on this diagram represents the location of a jet at 80° W on the cross section for that time; solid dots are used where the existence and location of the maxima are relatively certain, open circles are used where the existence or location of the jets are less certain. Connecting lines are drawn between the dots when it appears from the cross sections and upper air charts that the individual wind maximum would have moved continuously from one position to another if charts had been available at very frequent intervals. It is to be expected that in some situations the continuity of the wind maxima between two successive 12-hr periods is difficult to determine; no connecting lines have been entered on fig. 1 in these cases. The most striking fact shown on the graph is the predominance of a double wind maximum throughout the period. During some periods there are three or even four jets, but these periods are relatively short-lived. Another feature of the chart is a tendency for some of the systems to gradually move southward, ultimately disappearing at low latitudes. The three most regular of these are the jets which disappear on February 4, 12, and 23.

An examination of the 300- and 500-mb charts indicated that the appearance of new wind maxima at 80° W was often associated with the development of large disturbances to the west. Furthermore, a comparison of the movement and location of the jets with the frontal systems on the surface weather chart indicated that a close connection between the two often existed in the sense that each of the successive jets tended to be associated with an individual polar front in the vicinity of the Atlantic coast of North America. The remainder of this article will be devoted to a description of these phenomena as illustrated by several of the more clear-cut cases.

The period 21 Jan.—4 Feb.

This period begins with the appearance at 80° W of a new maximum at about 45° N. This maximum subsequently moves southward and disappears by the end of the period at about 30° N. Fig. 2 contains schematic diagrams showing the relative location of the surface frontal systems and the principal jets at various intervals during this 15-day period. On January 21st the newly formed weak wind maximum at 80° W is seen to be closely connected with the unusually well-developed "arctic front" in the middle of the continent.[1] The "older" jet to the south is seen to be associated with a weak frontal system on the surface. By January 23 the southern frontal system has become very weak on the surface weather chart, and the arctic front now contains the principal contrast between the cold continental air and the warm sub-tropical air — i. e. the arctic and polar fronts are merging into a new strong polar front. This is shown in

[1] The name "arctic front" is commonly used by American meteorologists to denote the (usually) shallow frontal system found in winter between the extremely cold surface air over the northern part of the continent and the more temperate polar air (often from a maritime source) to the south.

Fig. 2. Location of jet streams and surface frontal systems 21 Jan.—4 Febr. 1948. A jet stream is indicated by a hatched arrow, broken segments of the arrow indicating a diffuse jet. Conventional symbols are used to indicate the surface frontal positions.

the cross section for 1/23/0300 (fig. 3)[1]. This cross section shows definitely the existence of two wind maxima and shows particularly well the essential difference in the structure of the two systems. The southern jet, associated with a frontolysing surface frontal system, has its solenoidal field well developed only in the middle and upper troposphere, while the intensifying jet to the north has a greater portion of its solenoidal field in the lower troposphere.

The southern jet disappears at 80° W shortly after 1 23 0300 and on Jan. 28 there is only one jet stream with its associated front crossing the meridian. The jet and the front are very strongly developed, as shown in the corresponding cross section in fig. 3. The schematic chart in fig. 2 for 1 28/0300 suggests that the strong current entering the continent from the Gulf of Alaska will no longer flow around the trough in the southwestern United States but will cross the 80th meridian as a current distinct from that now crossing the longitude. This does occur and several days later a deep disturbance again develops in the western part of the United States (as shown on the schematic chart for 2/3/0300) with a new

[1] For brevity, a symbolism such as "1/23/0300" will be used to denote the date and time of a chart, the quoted figures referring to 0300 GCT on the 23rd of January.

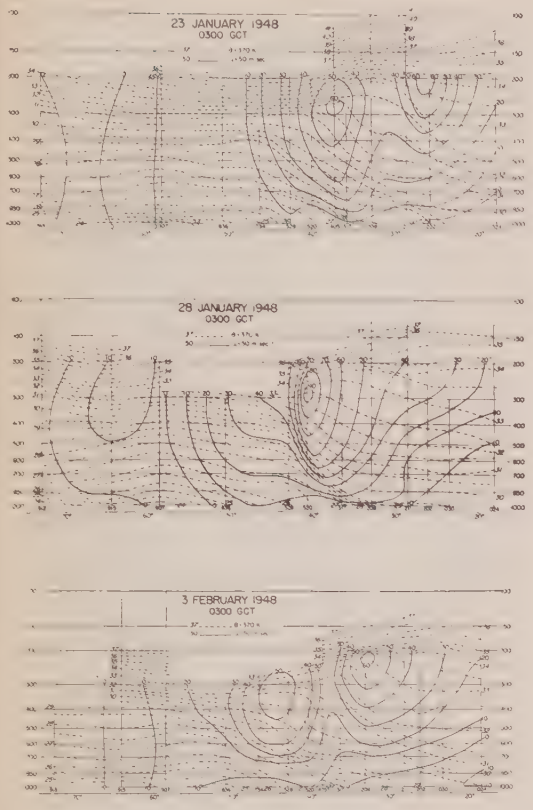


Fig. 3. Meridional cross sections at 80° W for 1/23/0300, 1/28/0300, and 2/3/0300 1948. Isentropes are entered in dashed lines, isolines of zonal wind speed in solid lines. Straight vertical lines show the location and height of the ærological soundings used in the analysis.

jet crossing the 80th meridian. In the meantime the jet and frontal system shown so well on the cross section for 1/28/0300 have moved southward and their intensity has decreased. However, a new Atlantic polar front is in the process of being formed by the frontal system associated with the new jet stream and a new cycle is beginning. In this light the general similarity of the cross sections and schematic charts for 1/23/0300 and 2/3/0300 is rather striking.

The synoptic events indicated in a rough manner by the schematic charts and cross sections in figs. 2 and 3, and their correspondence to the type of synoptic development frequently found in eastern North America, suggests that there is often an intimate relation between the formation and southward displacement of the jet at this fixed longitude and the southward displacement and subsidence of a mass of cold air in this region. The example described here took rather long (two weeks) to go through a "cycle". It is not believed that this case is necessarily representative in this respect; there are indications in the remainder of the period studied that the same process may occur at other times on a smaller and more rapid scale (and therefore more difficult to detect). It is also not suggested that the same sequence of appearances, displacements, and disappearances of a jet stream would be encountered at all longitudes in the same manner (e. g. a chart corresponding to fig. 1 for the meridian of 120° W would have been quite different from that for 80° W during this 49-day period).

The variations of potential temperature in the center of this jet are summarized in the top portion of fig. 4. There is a definite tendency for the current to be found at higher potential temperatures as it moves southward, although superimposed on the general trend are rather pronounced short-period fluctuations, the result of either errors in analysis or actual short-period changes in the jet at 80° W. If these latter are present, they are presumably connected with disturbances which are not sufficiently intense to appear as discontinuities

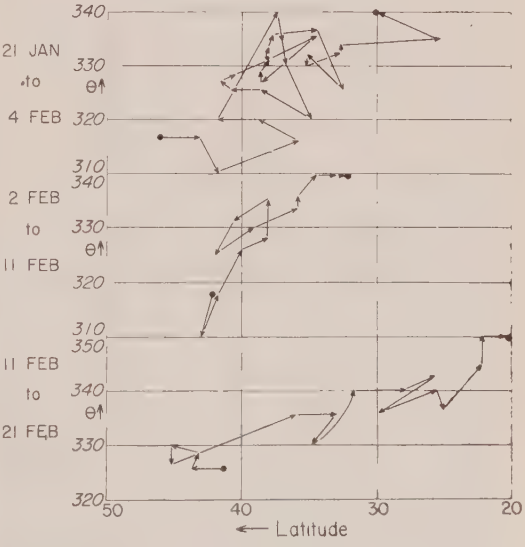


Fig. 4. Potential temperature (degrees Absolute) in the center of three jets at 80° W as a function of latitude of the individual jet. Each arrow represents either 12 or 24 hours.

of the jets on fig. 1. Some suggestion of the reality of these short-period fluctuations is contained in the first part of the recent article by Berggren, Bolin, and Rossby (1949).

The latitudinal shear (anticyclonic) of the zonal wind along the isentropic surfaces to

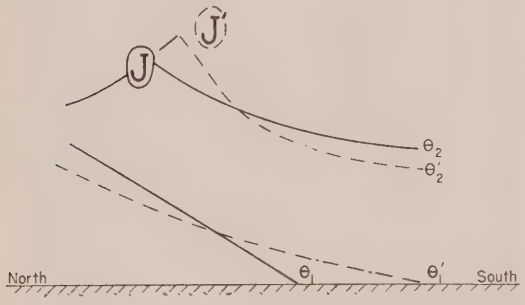


Fig. 5. Deformation of isentropic surfaces accompanying a shift of a jet to higher potential temperature.

the south of the jet in many of the analysed cross sections reaches a magnitude almost equal to that of the Coriolis parameter. Professor C.-G. Rossby has suggested to the writer that this may perhaps have something to do with the movement of the jet to higher elevations and potential temperatures. The process is indicated schematically in fig. 5. The near approach to inertia instability immediately to the south of and below the jet would presumably imply a greater freedom of meridional movement along the isentropic surfaces in this region than in the region farther to the south where the isentropic surfaces are more horizontal. If this meridional movement is primarily northward, the resulting net transport of air along the isentropic surfaces would tend to deform them in the manner indicated in fig. 5, and, *ceteris paribus*, cause a movement of the jet to a higher potential temperature. A study of the several cross sections reproduced in fig. 3 shows that the isentropic surfaces have been deformed in this manner (e. g. the surfaces $\Theta = 295$ and 310) but the importance of these meridional motions and their associated vertical motions in producing the observed isentropic deformation at a fixed longitude is impossible to decide until the effects of longitudinal (zonal) advection and non-adiabatic processes are eliminated. Unfortunately this is virtually impossible at the present time.

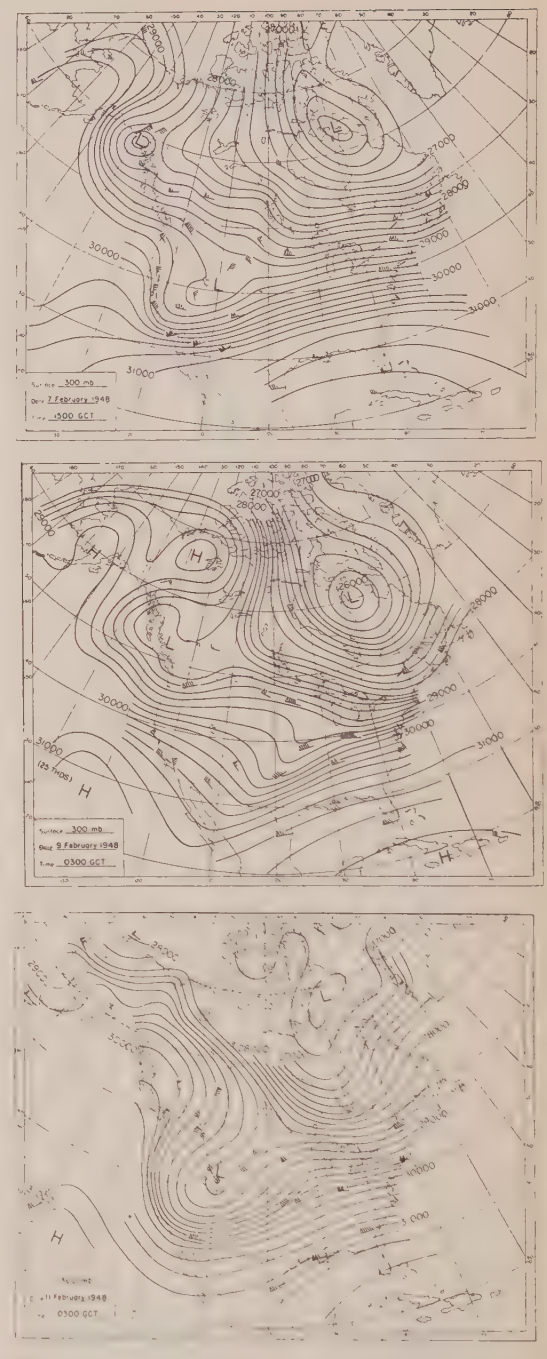
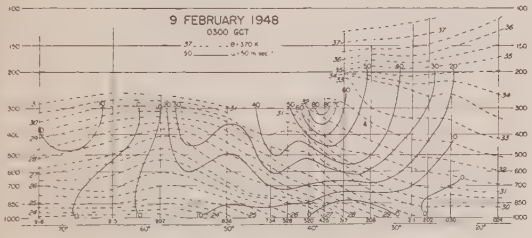


Fig. 6. 300-mb charts for 2/7/1500, 2/9/0300, and 2/11/0300 1948. Contours are for 200-ft intervals. Observed winds at this level are plotted with the following convention: solid triangular barbs = 50 mi hr^{-1}, full straight line barb = 10 mi hr^{-1}, half straight line barb = 5 mi hr^{-1}. Observed winds at a level 5000 ft lower are indicated by open triangular barbs.

The period 7—22 February

At the beginning of this period a new jet stream made its appearance at longitude 80° W at about 45° N (fig. 1) and subsequently moved southward and finally disappeared. However, the relation between the jet and the polar front on the surface was not as simple as in the case described above. The outbreaks of arctic air associated with the new jet occurred over the western part of the United States and therefore there was no replacement of the previous Atlantic polar front on the surface by a new frontal system at the beginning of the period. Also, the jet maintained its existence at a southerly latitude for some time after the surface frontal system had disappeared.

Fig. 6, containing the 300-mb charts for 2/7/1500, 2/9/0300, and 2/11/0300, shows rather clearly how the appearance of the new jet at 80° W was closely associated with the development and southward movement of a cyclonic vortex over the western part of the continent. In this type of development the cold arctic air in the lower levels is brought far to the south over the region of the Southwestern United States and occasionally even into California. This is apparently the reason



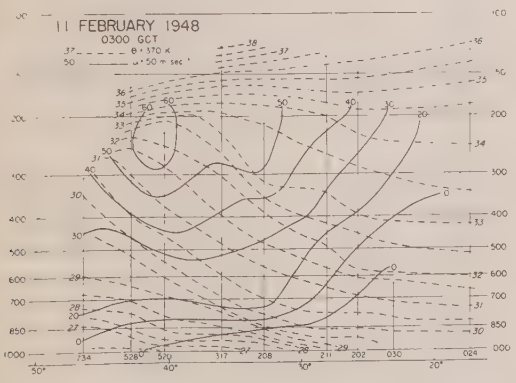
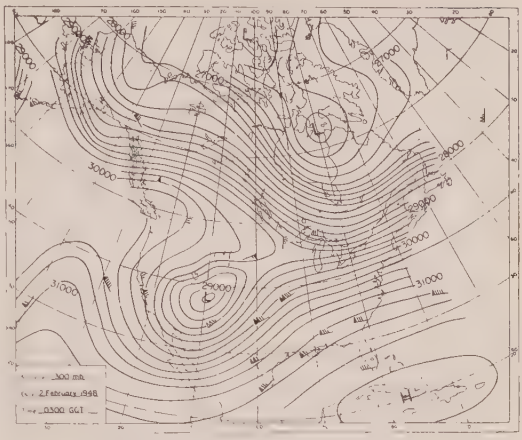


Fig. 7. Meridional cross sections at 80° W for 2/9/0300 and 2/11/0300, 1948. (See legend to fig. 3.)



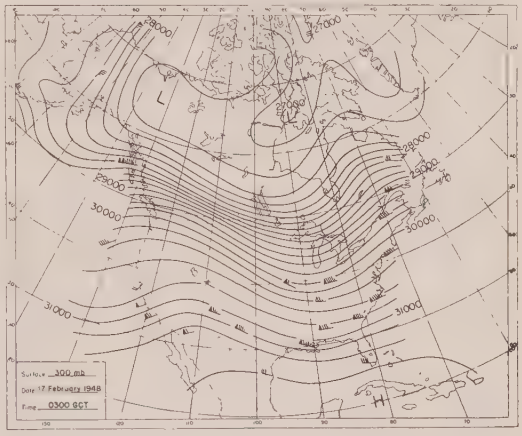


Fig. 8. 300-mb charts for 2/12/0300 and 2/17/0300, 1948. (See legend to fig. 6.)

why the behavior of the surface polar front over eastern North America did not correspond to that during the earlier period described above. The cross sections for the 9th and 11th of February (fig. 7) show the development of the new jet (very weak at 2/9/0300, 45° N — stronger at 2/11/0300, 40° N) and the weakening of the previous maximum (strongly developed at 2/9/0300, 39° N — barely discernible at 2/11/0300, 34° N).

The 300-mb charts for 2/9/0300 and 2/11/0300 seem to indicate that the new jet at 80° W is continuous with the strong current curving anticyclonically over Alaska. The 300-mb chart for 2/12/0300 (fig. 8) shows that this connection has been broken and the jet at 80° W will in the future be more connected

with what appears to be another current in the Pacific, south of the current in the Gulf of Alaska. Scattered wind and radiosonde reports from the Pacific (not reproduced) indirectly support this conclusion — which on the reproduced 300-mb charts for 2/12/0300 rests rather heavily on the single ship rawin west-southwest of San Francisco.

The cross section along the 80th meridian for 2/12/0300 (fig. 9) is interesting in that it is taken in a region of pronounced *confluence*, which according to NAMIAS and CLAPP (1949) is of some importance in the development of jet streams. In this respect it is interesting to note that the 300-mb chart for 2/12/0300 shows that the two jets exist upstream of the confluence region in as fully developed a state as they do in the confluence region.[1] Hydrostatically this requires a more-or-less longitudinally-constant cross stream temperature gradient in each current (in the middle and lower troposphere). Synoptically this in turn requires the southern jet in its southernmost position to flow around a cold low (trough) and the northern jet in its northernmost position to flow around a warm high (ridge). In the writer's opinion these conditions are usually found in cases of pronounced confluence. This suggests that the conclusion of NAMIAS and CLAPP that confluence necessarily causes an increase of the temperature gradient rests on a streamline-isotherm relationship which in many cases differs *in the essential features* from the streamline-isotherm relationship actually existing in regions of confluence. In other words, the process described by NAMIAS and CLAPP may actually occur in the atmosphere, but the conditions for its occurrence need not always exist in regions of streamline confluence.

The 300-mb chart for 2/17/0300 (fig. 8) shows that at this time there were still two maxima existing across the continent, in agreement with the indications of the chart for 2/12/0300. The existence of the southern current between 30° N and 35° N is also borne out by rawin reports at higher levels from stations in that latitude belt, these winds being stronger than those reported in the latitude belt 35—40° N. The cross section for 2/17/0300

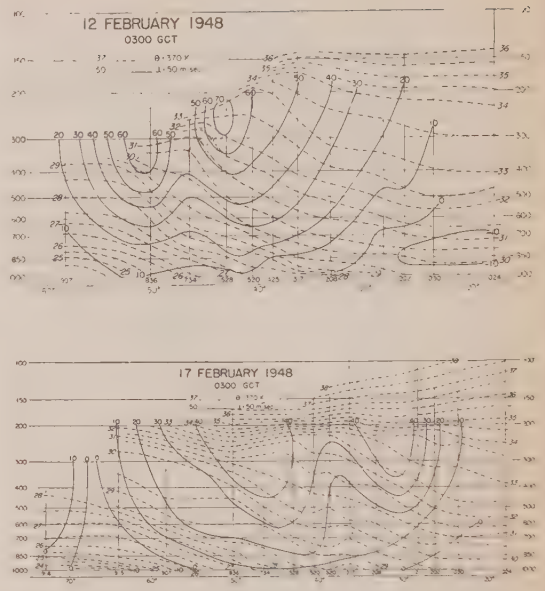


Fig. 9. Meridional cross sections at 80° W for 2/12/0300 and 2/17/0300, 1948. (See legend to fig. 3.)

(fig. 9) shows that the same changes have occurred in the structure of the southern jet since its first appearance at 80° W on the cross section for 2/9/0300 that occurred in the case of the jet studied above for the period 21 Jan. —4 Feb. However, the solenoid field in the lower troposphere associated with the southern jet on the cross section for 2/17/0300 seems to be somewhat stronger than the almost non-existent solenoid fields in the lower troposphere which accompanied the earlier jet in those latitudes (fig. 3). The surface frontal system which was associated with the jet as it developed in this case disappeared by the 15th of February.

The movement of the southern maximum between 2/17/0300 and 2/21/0300 was not as clear as it had been up until this time, but at 2/21/0300 there were still two maxima in existence, as shown by the 200-mb chart for that time (fig. 10) The general zonal orientation of the contours south of 30° N in this figure indicates, in an indirect and qualitative manner, that the current shown on that figure and on the cross sections in fig. 11 is not a local phenomenon but probably exists over a wide band of longitude. Unfortunately no observations were available to test this at other longitudes. However the 500-mb charts

[1] A computation of the mean gradient wind at the 300-mb level shows that in both currents the values are the same ($\pm$ 2 m sec ^{1}) before (over the western part of the continent) and at confluence (80°W).

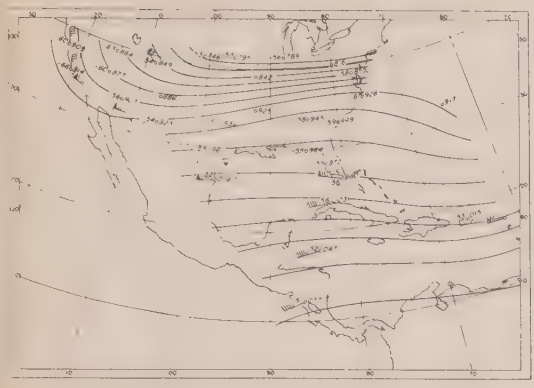


Fig. 10. 200-mb chart for 0300 GCT, 21 February 1948. Winds are plotted as stated in the legend to fig. 3 except that the winds for Havana (23° N, 82° W) and Swan Island (17° N, 84° W) are winds at 40 000 ft (~ 200 mb) and 35 000 ft (~ 250 mb) respectively, at 1500 GCT on the 21st (12 hrs after map time). Observed heights of the 200-mb surface are plotted in tens of feet with the first number (either 3 or 4) omitted, e. g. 003 = 40 030 ft. The observed height of 40 140 ft at Brownsville was considered incorrect. Temperatures are also plotted in °C; height reports unaccompanied by temperature reports are extrapolations made where the sounding terminated between 250 and 200 mb.

reproduced in the article by BERGGREN, BOLIN, and ROSSBY (1949), some of which are for this same period, indicate that a double maximum would also be found in the vicinity of the Greenwich meridian. Because of the low latitude, geostrophic wind computations are very

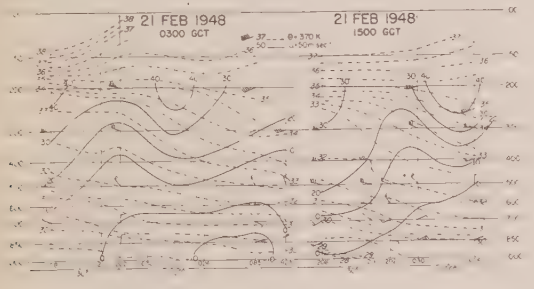


Fig. 11. Meridional cross sections at 80° W for 0300 and 1500 GCT, 21 February 1948. Winds are plotted in mi hr^{-1} (see legend to fig. 6). West winds are plotted horizontally as if coming from the North.

sensitive to small errors in the pressure field and therefore too much reliance cannot be placed on the numerical value of either the maximum speed in the center of the jet of fig. 11 or on the horizontal shear indicated by the isolines of u. To demonstrate that the existence of the current on the maps and cross sections does not depend on the existence of a systematic error in the observed pressure field, observed pibal and rawin reports have been included on the cross sections for 2/21/0300 and 2/21/1500 in fig. 11. The geostrophic (gradient) and observed winds are in fair agreement except for the last wind observation reported above Managua (085) at 2/21/0300. No trace of this southern current could be found after 2/22/1500. The continued existence of this jet for a week after the surface frontal system had disappeared (in contrast to the case of Jan. 21—Feb. 4) agrees with the observation that this jet was reinforced, so to speak, by the existence in the Pacific of a strong current at low latitudes.

The variation of the potential temperature in this jet is summarized in the lower portion of fig. 4, which shows the same type of variation of potential temperature as did the jet of 21 Jan.—4 Feb. A similiar graph for the jet which moved southward during the period 2 Feb.—12 Feb. shows the same changes.

It appears likely from the time scale indicated by fig. 1 that the phenomena represented thereon are not identical with those reported by CRESSMAN (1950). In the writer's opinion the southward displacement of hemispheric jets reported by CRESSMAN to occur during some periods must in some fashion be the statistical result of the processes shown in fig. 1 occurring at several longitudes around the hemisphere more-or-less simultaneously.

Acknowledgment

The writer wishes to especially acknowledge the large contribution made by Dr H. L. KUO with his participation in the primary phases of this project.

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A Re-examination of the "Leipzig Wind Profile" Considering some Relations between Wind and Turbulence in the Frictional Layer

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Abstract

The "Leipzig Wind Profile" is a unique example of a representative wind distribution in the frictional layer which resulted from MILDNER's set of 28 pilot-balloon observations with two theodolites on October 20, 1931, near Leipzig for a stable weather situation. It is shown, that the vectors of geostrophic motion and ground drag follow from the wind profile when a scalar austausch is assumed. The re-computed vertical austausch-distribution indicates that the energy for maintaining the turbulence in a steady frictional wind profile is taken from the potential energy of the horizontal pressure field by means of the steady flow of air across the isobars.

The frictional layer is the atmospheric layer adjacent to the ground where turbulent stresses are significant. Consequently, the frictional layer is characterized by considerable deviations of the observed wind from the theoretical air motion which follows from the pressure distribution with the aid of the equations of motion when friction is neglected.

Let us assume that the wind in any given horizontal plane is a steady uniform rectilinear flow. Then, the variations of the turbulent stresses in the horizontal directions are negligible such that only the vertical variations need be considered. The equation of motion for this case is

$$\mathbf{T}' = \varrho f i (\mathbf{v} - \mathring{\mathbf{v}}) \qquad (1)$$

when the prime denotes the partial differentiation with respect to height z (meters) and

$\mathbf{T}$ = horizontal vector of turbulent stress, dynes/cm²;
$\mathbf{v}$ = horizontal wind vector, m/sec;
$\mathring{\mathbf{v}}$ = horizontal geostrophic wind vector, m/sec;
f = Coriolis parameter = 0.0001458 sin Φ, sec^{-1};
Φ = geographic latitude, degrees;
ϱ = density of the air, g/cm³;
$i = \sqrt{-1}$.

A vector notation is used where the unit vector in the horizontal plane is $\mathbf{i}$ when $i\,\mathbf{i}$ denotes the horizontal direction perpendicular to $\mathbf{i}$. The positive $i\,\mathbf{i}$ direction is defined in fig. 1.

Eq. (1) shows that—for steady states—the internal friction ($\mathbf{T}'$) balances the pressure gradient ($\nabla p = -\varrho f i \mathring{\mathbf{v}}$) and the Coriolis force ($\varrho f i \mathbf{v}$); the physical units of the terms in eq. (1) are dynes/cm³.

Let us define the turbulent stress as proportional to the wind shear, i.e.

$$\mathbf{T} = A \mathbf{v}' \qquad (2)$$

where A = austausch-coefficient, g/cm sec; A is invariably positive and a function of z. In general, a linear relation between two vectors—as in eq. (2)—defines a tensor. However, it will be shown below that a scalar austausch coefficient is sufficient and—owing to our present knowledge of wind distributions—necessary to describe the characteristics of turbulence in the frictional layer.

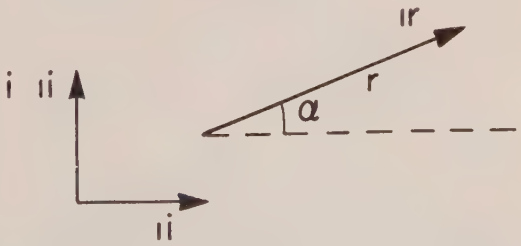

Fig. 1. The unit vectors and an arbitrary vector in a horizontal plane.

Let r and α be the absolute value and the azimuth of an arbitrary horizontal vector $\mathbf{r}$; see fig. 1. Then

$$\mathbf{r} = \mathbf{i}\, r\, e^{i\alpha} = \mathbf{i}\, r \cos\alpha + i\,\mathbf{i}\, r \sin\alpha \qquad (3)$$

The differentiation of eq. (3) with respect to z yields:

$$\mathbf{r}' = \mathbf{r}\left(\frac{r'}{r} + i\,\alpha'\right) \qquad (4)$$

Since the scale product of two vectors which are perpendicular to each other equals zero, we obtain from eq. (4) the following geometrical relations:

$$\mathbf{r} \cdot i\,\mathbf{r} = \mathbf{r}' \cdot i\,\mathbf{r}' = 0 \qquad (5\,a)$$

$$\mathbf{r}' \cdot i\,\mathbf{r} = -r^2\,\alpha' \qquad (5\,b)$$

Upon multiplying eqs. (1) and (2) by $i\,\mathbf{v}$ and $i\,\mathbf{v}'$, respectively, when eq. (5 a) is considered, we obtain

$$i\,\mathbf{v} \cdot \mathbf{T}' = -\varrho f\,\mathbf{v} \cdot (\mathbf{v} - \mathbf{v}) \qquad (6)$$

$$i\,\mathbf{v}' \cdot \mathbf{T} = i\,\mathbf{v}' \cdot \mathbf{v}'\,A = o \qquad (7)$$

thus:

$$i\,\mathbf{v} \cdot \mathbf{T}' = (i\,\mathbf{v} \cdot \mathbf{T})' = -\varrho f\,\mathbf{v} \cdot (\mathbf{v} - \mathring{\mathbf{v}}) = \\ = -\varrho f\,(v^2 - v\mathring{v}\cos\alpha). \qquad (8)$$

where v and $\mathring{v}$ are the absolute values of $\mathbf{v}$ and $\mathring{\mathbf{v}}$ when α is the angle between $\mathbf{v}$ and $\mathring{\mathbf{v}}$. Let T be the absolute value of $\mathbf{T}$ and let the subscript o denote the boundary values of v, T and α at the ground level $z = 0$. The boundary conditions of the problem are

$$v_o = 0,\ T_o \neq 0,\ 0 < \alpha_o < \frac{\pi}{2}; \qquad (9)$$

with the aid of eqs. (5 b) and (9) it follows from eq. (8) that

$$A = \frac{1}{v^2\alpha'} \int_0^z \varrho f(v^2 - v\,\mathring{v}\cos\alpha)\,dz \qquad (10)$$

Eq. (10) can be used for the direct determination of A at any level of the frictional layer provided v, α and $\mathring{v}$ are known with sufficient accuracy. Eq. (10) corresponds to the formulas of SOLBERG (see V. BJERKNES and collaborators, 1933) and FJELSTAD (1929, see also SVERDRUP, 1933).

Wind observations, which are accurate and representative enough in order to permit the application of eq. (10) are extremely scarce. MILDNER'S (1932) set of 28 pilot-balloon observations with two theodolites on October 20, 1931, near Leipzig, appears to be the only one which is reliable with regard to the assumptions of steady and uniform rectilinear flow and the elimination of random turbulent fluctuations.

By means of a graphical differentiation as based on a smoothed line connecting the endpoints of the averaged observed v-vectors in the frictional layer, MILDNER obtained with the aid of SOLBERG'S formula and the geostrophic wind from the weather maps the following A-values (see Table 1).

Table 1. MILDNER'S austausch-distribution.

z, m.....	80	125	190	240	295	405	460	510
A, g/cm sec.	125	270	310	500	246	117	70	70

For the sake of brevity let us refer to Mildner's wind observations as the "Leipzig Wind Profile."

In re-examining the "Leipzig Wind Profile" this author found that the results which follow from eq. (10) are considerably affected by relatively small variations of the azimuth of the geostrophic wind. It had to be concluded that the geostrophic wind as taken from the weather map might be too inaccurate in comparison with the value derived from the wind profile when certain geometrical relations are considered.

Let the subscripts x and y denote the horizontal vector components in the directions of $\mathbf{i}$ and $i\,\mathbf{i}$, respectively. For convenience, let us assume that $\mathring{\mathbf{v}}' = 0$ and let the direction of $\mathbf{i}$ be parallel to $\mathring{\mathbf{v}}$. The assumption $\mathring{\mathbf{v}}' = 0$ is generally a satisfactory one; it means that the horizontal density distribution is uniform or thermal winds are negligible.

Moreover, let us assume that in the frictional layer the terms $\varrho'\,\mathbf{v}$ are negligible in comparison with $\varrho\,\mathbf{v}'$. Considering the relative shallowness of the atmospheric layer in question ($0 \leqq z \leqq 1{,}000$ m), $\varrho' = 0$ appears to be a tolerable approximation.

There exist the two significant levels z_1 and z_2 in the frictional layer. At $z = z_1$, v_y, the wind component perpendicular to $\mathring{\mathbf{v}}$, reaches its first extreme value when at $z = z_2$, v_x, the wind component parallel to $\mathring{\mathbf{v}}$, reaches its first extreme value. Owing to the geometrical shape of the wind profile and the boundary

conditions $v_{x0} = v_{y0} = 0$ at $z = 0$, these extreme values are maximum values.

Eq. (1) yields, with the aid of eqs. (9) and the above definitions and assumptions:

$$T_x = T_{0x} - \varrho f \int_0^{\tilde{z}} v_y \, dz = A \, v'_x \quad (11a)$$

$$T_y = T_{0y} + \varrho f \int_0^{\tilde{z}} (v_x - \mathring{v}) \, dz = A \, v'_y \quad (11b)$$

Since $A > 0$ at any level, when $v_x' = 0$ at $z = z_2$ and $v_y' = 0$ at $z = z_1$, we must have

$$T_{0x} = \varrho f \int_0^{\tilde{z}_2} v_y \, dz \quad (12a)$$

$$T_{0y} = -\varrho f \int_0^{\tilde{z}_1} (v_x - \mathring{v}) \, dz \quad (12b)$$

The direction of $\mathbf{T}_0$ follows from the direction of the wind near the ground. Unknown values are the correct amounts of α_0 and $\mathring{v}$, i.e. the azimuth (relative to $\mathbf{T}_0$) and the value of the geostrophic wind.

For evaluating $\mathring{\mathbf{v}}$ from the "Leipzig Wind Profile", the following method was used: A tentative α_0 is assumed; then, the observations yield v_x, v_y, z_1 and z_2; from eq. (12a), T_{0x} can be computed which yields T_{0y} by means of the general relation:

$$T_{0y} = T_{0x} \tan \alpha_0; \quad (13)$$

with the aid of eq. (12b) one obtains:

$$\mathring{v} = \frac{1}{z_1} \left\{ \int_0^{z_1} v_x \, dz + \tan \alpha_0 \int_0^{z_2} v_y \, dz \right\} \quad (14)$$

Four different values of α_0 were used in re-examining the "Leipzig Wind Profile" and resulted in different values of z_1, z_2, T_0 and $\mathring{v}$ (see Table 2). All computations were carried out for successive 50 m-levels with the aid of graphical-numerical integrations.

Table 2. Characteristics of the "Leipzig Wind Profile" for different α_0.

α_0	23.4	24.9	26.1	27.0	degrees
z_1	230	240	235	250	m
z_2	875	900	880	900	m
T_0	4.23	4.86	5.31	5.77	dynes/cm²
$\mathring{v}$	15.77	16.66	17.51	17.95	m/sec
$\{T_y/v'_y - T_x/v'_x\}_m$	—18	—10	1	7	g/cm sec

In order to decide which α_0 is correct, a trial and error method was applied by computing the austausch-coefficient A with the aid of eq. (11a) and eq. (11b) from $A = T_x/v'_x$ and $A = T_y/v'_y$. Since A was defined as a scalar height function, both of these equations should result in the same A-values at any level.

The lowest line of Table 2 shows height-averages between 50 and 400 m of $T_y/v'_y - T_x/v'_x$, which would correspond to the difference of fictitious tensor components when $A_{xx} = T_x/v'_x$ and $A_{yy} = T_y/v'_y$. When A is a scalar, we should expect these differences to equal zero. It follows from Table 2 that the computation based on $\alpha_0 = 26.1°$ yields a satisfactory fit. It is rather obvious that abandoning the condition of a scalar austausch must result in arbitrary families of (A_{xx}, A_{yy})-distributions unless T_0 and $\mathring{v}$ are fixed accurately by independent observations. For unique results, it is necessary to assume $A_{xx} = A_{yy}$, i.e. a scalar A. Consequently, the outer conditions of the "Leipzig Wind Profile" can be listed as follows:

$\mathring{v} = 17.51$ m/sec
$T_0 = 5.31$ dynes/cm²
$\alpha_0 = 26.1°$
$\varrho = 0.00125$ g/cm³, $f = 0.000140$ sec⁻¹ $\Big\} \varrho f = 1.425\,10^{-7}$ g/cm³sec

The formulation of the geophysical conditions of wind profile observations is incomplete when data about the hydrostatic stratification of the air and the characteristics of the roughness of the earth's surface is missing. MILDNER (1932) stated that the observations were taken in a uniform "warm" air mass and that no indication of convectional processes occurred from 0915—1615 hours during which period the 28 pibal ascents were carried out.

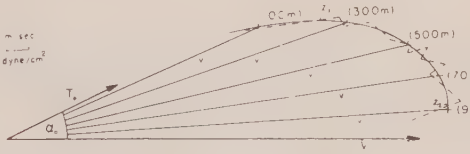

Fig. 2. The representative hodograph of the "Leipzig Wind Profile" from 50—900 meters. The broken line indicates MILDNER's averaged observations.

In the same air mass, the 1400 hour sounding at the aerological station Lindenberg (less than 100 miles from Leipzig) showed a rather uniform lapse rate of $-0.65°$ C/100 m in the layer under consideration, which corresponds to an increase of potential temperature of 0.35° C/100 m.

The characteristics of the ground can be described by the fact that the ascents started over a grass-covered airfield while the surroundings were rather plain. MILDNER states that the wind passed over the city before the air arrived at the station which may have influenced the turbulence to an unknown degree.

MILDNER's examination of the weather maps resulted in a 2.5 mb/100 km horizontal pressure gradient which was independent of height in the lower 1 km layer; thus, $\mathring{v} = |\nabla p| / \varrho f = 17.5$ m/sec which equals our value in the above list.

The "Leipzig Wind Profile" is a rather unique and—at the present time—unparalleled example of a representative wind distribution in the frictional layer of the atmosphere. Therefore, and in order to stimulate further research, it appears justifiable to publish in Table 3 the representative wind data as elaborated from Mildner's graph. In the process of adjustment v_x, v_y, v_x' and v_y' were considered to be smooth height functions. The differences between original and adjusted or, representative wind components as illustrated by the hodograph (fig. 2) correspond to a standard deviation of approximately $\pm$ 0.05 m/sec; the small magnitude appears to be due to both the stable weather situation and the large number of observations.

From eqs. (11 a, b) it follows that in a scalar austausch theory

$$A = \sqrt{T_x^2 + T_y^2} \,/\, \sqrt{v_x'^2 + v_y'^2} = T/v'. \quad (15)$$

Another and previously unknown formula for the austausch-coefficient can be derived as follows. Regarding $\mathring{\mathbf{v}}' = 0$, eq. (2) can be written

$$\mathbf{T} = A\ (\mathbf{v} - \mathring{\mathbf{v}})' \quad (16)$$

Upon multiplying eqs. (1) and (16) by $i\ (\mathbf{v} - \mathring{\mathbf{v}})$ and $i\ (\mathbf{v} - \mathring{\mathbf{v}})'$, respectively, one obtains, when eq. (5 a) is considered,

$$i\ (\mathbf{v} - \mathring{\mathbf{v}})' \cdot \mathbf{T} = (i\,(\mathbf{v} - \mathring{\mathbf{v}}) \cdot \mathbf{T})' = -\varrho f (\mathbf{v} - \mathring{\mathbf{v}})^2 \quad (17)$$

Let s be the absolute value of $(\mathbf{v} - \mathring{\mathbf{v}})$ when β is the angle between $(\mathbf{v} - \mathring{\mathbf{v}})$ and $-\mathring{\mathbf{v}}$; then, $\tan \beta = -v_y/(v_x - \mathring{v})$ and

$$\mathbf{v} - \mathring{\mathbf{v}} = -\mathbf{i}\ s\ e^{-i\beta};$$

$$(\mathbf{v} - \mathring{\mathbf{v}})' = (\mathbf{v} - \mathring{\mathbf{v}}) \left(\frac{s'}{s} - i\,\beta' \right) \quad (18)$$

With the aid of eqs. (17), (18) and the boundary conditions (9) considering that $(i\,(\mathbf{v} - \mathring{\mathbf{v}}) \cdot \mathbf{T})_0 = -i\mathring{\mathbf{v}} \cdot \mathbf{T}_0 = -\mathring{v}\, T_0 \sin \alpha_0$, we obtain finally

$$A = \left(T_0 \mathring{v} \sin \alpha_0 - \varrho f \int_0^z s^2\, dz \right) \Big/ s^2 \beta' \quad (19)$$

The three different expressions for the austausch-coefficient, eqs. (10), (15) and (19), result in A-values which are listed in Table 3. In view of certain numerical integration difficulties which arise from insufficient knowledge of wind distribution in the lowest 50 m-layer, the differences of the three A-series appear tolerable. The arithmetic average of the three A-distributions, therefore, may be called the "representative average" of the austausch-distribution from the "Leipzig Wind Profile". Table 3 shows also the standard deviation of the average. With reference to Table 1, MILDNER's A-values are of the true order of magnitude; however, they are too large below 300 m and too small above owing to incorrect assumptions with regard to the azimuth of the geostrophic wind.

It appears significant that the vertical distributions of A and v_y are rather similar to each other (see Table 3). This may indicate that the energy for maintaining austausch or turbulence in a steady frictional wind profile is taken from the potential energy of the horizontal pressure distribution by means of v_y, the steady flow from higher to lower pressure which, naturally, must reduce, and finally, eliminate the horizontal pressure differences or, the geostrophic motions, when these are not renewed by other atmospheric processes.

Fig. 2 shows the "Leipzig Wind Profile" as an illustration of an observed frictional wind spiral. Fig. 3 shows the corresponding stress spiral.

Finally, it may be noted that the direct integration of eq. (1) with regard to the boundary conditions at $z = \infty$: $T_\infty = 0$, $\mathbf{v}_\infty = \mathring{\mathbf{v}}$, yields

Table 3. The "Leipzig Wind Profile" and the subsequent austausch-distribution. z = height (m); v_x = wind component parallel to the geostrophic wind (m/sec); v_y = wind component perpendicular to the geostrophic wind (m/sec); A = austausch-coefficients (g/cm sec).

z	v_x	v_y	A	A	A	A	
			eq. (15)	eq. (10)	eq. (19)	representative average	standard deviation
0	0.00	0.00	—	—	—	—	—
50	9.15	4.35	152	116	127	132	± 8
100	10.45	4.64	173	138	177	163	12
150	11.58	4.80	179	158	182	173	8
200	12.60	4.95	183	170	184	179	5
250	13.48	4.96	184	174	183	180	3
300	14.30	4.90	181	176	181	179	1
350	14.97	4.78	176	173	175	175	1
400	15.62	4.60	167	166	164	166	1
450	16.28	4.29	159	160	160	160	0
500	16.83	4.00	149	151	151	150	1
550	17.30	3.71	141	145	141	142	1
600	17.70	3.37	136	140	135	137	1
650	17.99	3.07	129	135	120	128	4
700	18.23	2.73	119	128	107	118	6
750	18.42	2.43	113	119	92	108	8
800	18.60	2.06	100	105	80	95	8
850	18.66	1.70	83	89	70	81	6
900	18.68	1.31	73	76	60	70	5
950	18.62	0.91	57	67	50	58	5

$$T_{ox} = \int_0^\infty \varrho f v_y \, dz \qquad (20\,a)$$

$$T_{oy} = -\int_0^\infty \varrho f (v_x - \mathring{v}) \, dz \qquad (20\,b)$$

These relations were used previously (Rossby and Montgomery, 1935). However, they appear less practicable than eqs. (12 a) and (12 b) since the latter can be applied for a final and well-defined height interval in which both assumptions $\mathring{\mathbf{v}}' = 0$ and $\varrho' = 0$ appear more tolerable than for layers of undetermined height.

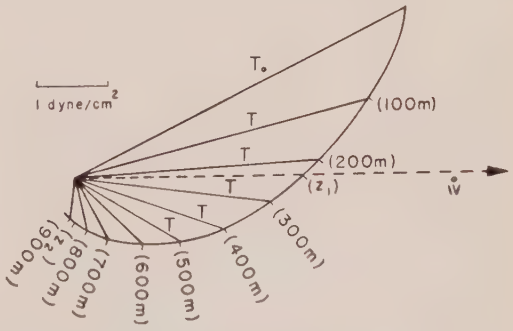

Fig. 3. The representative stress spiral of the "Leipzig Wind Profile". The significant levels z_1 and z_2 where the stress vector is parallel and perpendicular to the geostrophic wind vector are indicated in Figures 2 and 3.

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SHORTER CONTRIBUTION

The Mean Wind and Temperature Distribution through a Flat Upper Ridge in Winter

By YI-PING HSIEH[1], University of Chicago[2]

(Manuscript received 7 March 1950)

Abstract

A mean cross section is constructed from a number of individually analyzed daily cross sections through flat upper ridges, using the position of the maximum of the geostrophic zonal wind at upper levels as reference for the meridional coordinate. Computed wind velocity, wind shear, and vorticity distribution are discussed. A comparison of the results with the meridional cross section of PALMÉN and NEWTON is emphasized.

Introduction

Since PALMÉN and NEWTON (1948) published their work about the mean wind and temperature distribution in the vicinity of polar front, it has been widely referred to and used in the consideration of problems in connection with the general circulation of the atmosphere, although the authors did not claim that their result represented the general situation of the atmosphere. One of the purposes of their work was to study the structure of the polar front and its relation with the wind and temperature fields. Therefore, they selected cases with a well-defined polar front. As a result, the intensity of the wind and temperature fields shown in their cross section can represent only the situation in a meridional cross section through a well-defined front. Furthermore, the individual cases they selected to construct their mean picture are so chosen that the cross sections are near the mean trough at the east coast of North America (although cases of strong curvature were not used). For these reasons one might expect that the wind and temperature fields discussed by PALMÉN and NEWTON are more intense than those found at other longitudes and times.

In order to clarify our ideas about the distribution of wind and vorticity in the meridional cross section, the preparation of a mean cross section at the same longitude of 80° W through a 'flat' ridge would thus seem to be advisable. 'Flat' ridges are desirable in order that the effects on the gradient wind speed of the trajectory curvature be held to a minimum. The same longitude of 80° W is desirable because the location of aerological stations along this meridian is almost ideal.

An additional consideration in selecting these cases was the elimination of those where more than one jet was well developed. The nine cases are those at 0300 GCT on the following dates: 30 November and 2 December, 1948; 17 January, 10 February, 27 February, 21 March, 25 March, 7 December, and 17 December, 1949. Fig. 1 is the mean 500 mb

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[2] Published as a contribution of a research on the general circulation of the atmosphere, sponsored by the office of Naval Research.

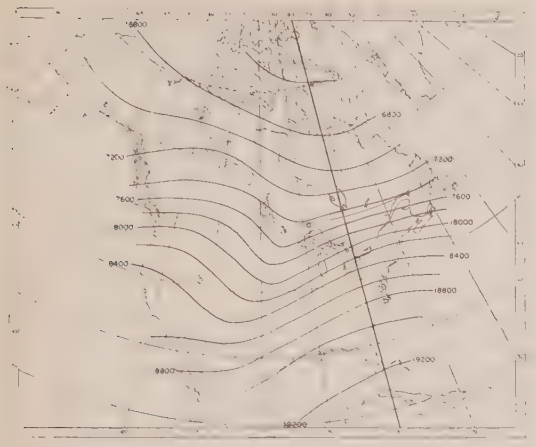

Fig. 1. Mean 500-mb surface of nine cases in winters of 1948 and 1949. Straight heavy line indicates the position of the mean cross section. Contours are drawn for 200-ft. intervals.

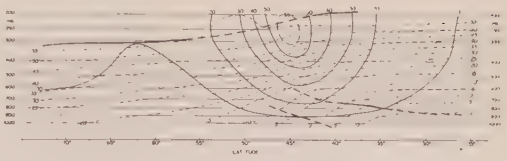

Fig. 2. Mean temperature and zonal component of geostrophic wind. Heavy lines indicate mean positions of frontal boundaries and tropopause. The dashed lines are isotherms (°C, slanting numbers) and the solid lines are lines of equal geostrophic zonal wind (m sec^{-1}, upright numbers).

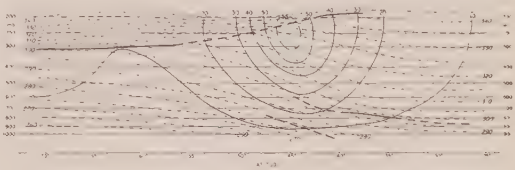

Fig. 3. Mean potential temperature (°K, dashed lines) and geostrophic zonal wind (m sec^{-1}, solid lines) for the same cross section shown in fig. 2.

map for the 9 cases and gives a general impression of the situation. The flat ridge along the meridian of 80° W extends from the tropics up to about 55° N.

The mean cross section was determined in the same fashion as was that of PALMÉN and NEWTON, except for one difference—the position of the upper maximum wind (e.g. at 300 mb) is taken as the moving reference for the meridional coordinate instead of the polar-front layer. This was necessary because the variation of the slope of the polar front in the ridge region turned out to be greater than in the situation covered by the above authors.

Temperature and Wind Fields

The mean temperature and wind fields are shown in fig. 2[1] and potential temperature is shown in fig. 3. The mean front is not as distinct as in PALMÉN's and NEWTON's cross section, and its slope is smaller. Since the polar-front layer is not taken as the coordinate reference, the process of averaging may obliterate some of the frontal characteristics. Also, as generally recognized, the polar front is less distinct and has a smaller slope near the ridge than near the trough. Thus, in fig. 2 the polar front shows up as a broad transitional zone, and is quite indistinct near the surface.

Just as in their cross section, the baroclinity is not limited to the frontal zone, but exists in the cold and warm air as well. The baroclinity in the warm air decreases rapidly towards the south so that almost barotropic conditions exist south of 35° N. However, the baroclinity in the cold air extends to high latitudes.

Although the details of the tropopause zone are damped out in the mean cross section, a gentle slope upwards from north to south is indicated, with a maximum slope above the northern extension of the polar front. It differs somewhat in this respect from PALMÉN and NEWTON's result, which shows the tropopause to reach its lowest elevation just above the northern extension of the polar front. The individual cases which make up the mean cross section show the same result. The jet stream shows a maximum wind of 55 m sec^{-1} at the 250-mb level. Since our front does not reach the 500-mb surface, nothing can be said about the relative positions of the jet stream at upper levels and the intersection of the polar front with the 500-mb surface. However, the upper jet is approximately above a temperature of —25° C at 500 mb, which agrees, in general, with the statement by PALMÉN and NEWTON that the jet is located above the intersection of the 500-mb surface and the polar front.

The location of the maximum wind velocity at approximately 45° N, the same as that of PALMÉN and NEWTON, must be considered as a coincidence. As they state, in their cases the jet was situated somewhat farther to the north than usual, and this fact is probably the explanation for the agreement in

[1] The references to latitude used hereafter are not intended to imply the actual mean latitudes of the elements referred. Latitudes are used merely for brevity in description, as done by PALMÉN and NEWTON.

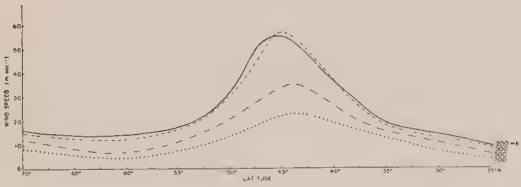

Fig. 4. Geostrophic wind profiles at various isobaric surfaces.

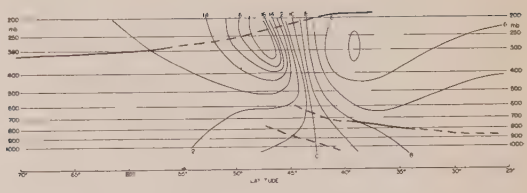

Fig. 6. Absolute vorticity (10^{-5} sec^{-1}).

latitude. It is, of course, evident from these two mean cross sections one cannot conclude that the maximum westerly wind will appear at 45° N in other meridional cross sections (for example, along the meridian of 5° W).

The wind distribution at different isobaric surfaces is shown in fig. 4. The position of maximum wind shifts toward north as height increases.

Wind Shear and Vorticity

Fig. 5 shows the values of horizontal wind shear for the same cross section. Strong shear appears on both sides of the jet, cyclonic to its north and anticyclonic to its south, although both the intensity and the concentration of the shear pattern are less than that shown by PALMÉN and NEWTON. In the past few years, the strong anticyclonic shear to the south of the jet has attracted much attention, particularly with respect to the instability of the horizontal flow in this region. KUO (1949) in a recent paper argued that if such strong shear happened only in a limited region, its effect should be damped by the influence of the much larger neighboring region with weak shear.

In our cross section, the absolute value of the shear of the computed geostrophic wind south of the jet is less than that of the coriolis parameter. As all the cases we have chosen are through a very weak ridge, and the computed geostrophic wind speed fairly agrees with the observed, the influence of the curvature may be considered as small. It seems reasonable then that south of the jet the actual value of the absolute vorticity in this case remains greater than zero. Therefore, the horizontal flow pattern is stable by the consideration of conservation of angular momentum.

Fig. 6 gives the distribution of the absolute vorticity in the cross section. Since we have considered the curvature small, the absolute vorticity is taken as the sum of the shear and the coriolis parameter. There is a maximum value of vorticity of 17×10^{-5} sec^{-1} to the north of the jet stream, and a minimum value of 5×10^{-5} sec^{-1} to its south. The contrast between the maximum and the minimum values of the absolute vorticity is not as distinct as in the cross section shown by PALMÉN and NEWTON.

The distribution of absolute vorticity at different isobaric surfaces is presented in fig. 7. It shows clearly that the absolute vorticity is fairly constant to the south and north of the jet stream, and that on the south side this constant value is greater than zero. It remains almost constant at the southern part of the cross section. At the northern end it increases very slowly and approaches the polar vorticity of earth rotation.

Fig. 7 suggests that there exist three relatively distinct zones, each with characteristic type of vorticity distribution, a northern zone of almost uniformly high vorticity, a transition zone in which the vorticity decreases from a high to a low value, and a southern zone of almost uniformly low vorticity.

The idea of zonal distribution of vorticity has been used by PLATZMAN (1949) in a theoretical study of upper tropospheric disturbances. He based his con-

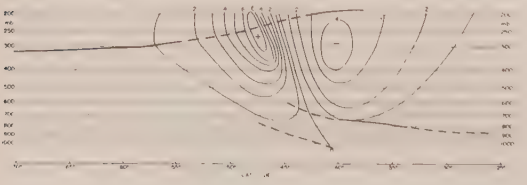

Fig. 5. Horizontal wind shear (10^{-5} sec^{-1}). Positive numbers indicate cyclonic shear, negative numbers anticyclonic shear.

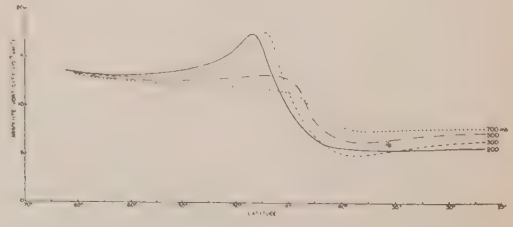

Fig. 7. Absolute vorticity profiles at isobaric surfaces.

sideration on PALMÉN and NEWTON's data and used a simplified model. ROSSBY (1947) has pointed out that a regime of uniform absolute vorticity within an extensive polar cap would tend to be established through the operation of large-scale lateral mixing. Fig. 7 indicates that there might be several zones with dominant large-scale lateral mixing processes and fairly constant vorticity separated by transitional zones with intensive meridional circulation and sharp change of vorticity.

Some Additional Remarks

The general characteristics of the mean distribution of the wind velocity, wind shear, and vorticity in the cross section through the flat ridges agree, in general, with that near troughs as shown by PALMÉN and NEWTON, but both the intensities and the contrast between the extreme values are less distinct than in the latter cross section. This is the situation when the geostrophic wind relationship is assumed to hold. If we consider the influence of curvature, even if we take a large radius of curvature, for instance 2,500 km (as the contours of the upper isobaric surfaces indicate), the situation may be reversed because of the great influence of the curvature on the wind speed for anticyclonic flow. A comparison of the geostrophic wind with the observed wind in the individual cases shows that they agree fairly well, although the computed geostrophic wind is usually about 5—10 per cent stronger than the observed. If we apply the gradient wind equation by using the curvature of the streamline, the result will give ridiculously high values for the computed wind speed in the high levels. It seems, therefore, that although the contours of the isobaric surfaces in our cases show some weak anticyclonic curvature, the trajectories are nearly straight because of the movement of the system, and furthermore that the curvature influence is probably compensated by the non-gradient flow.

Acknowledgements

The author wishes to express his appreciation for the benefit of many fruitful discussions with Profs. C. G. ROSSBY and E. PALMÉN, and Mr. C. W. NEWTON, and his gratitude to Mr. N. A. PHILLIPS for reading the manuscript.

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Letter to the Editor

On the Role of the Tropics in the General Circulation of the Atmosphere

Dear Sir:

In a recent issue of Tellus (Vol. 2, No 1) Mr Herbert Riehl published an interesting article on the subject indicated above, bringing new results and valuable information concerning the structure of the tropical atmosphere. My own meteorological and aerological observations in Mogadiscio (Italian Somaliland, Lat. 2°02′03″ N, Long. 45°21′20″ E) during the International Polar Year 1932—1933 appear to have considerable bearing on the problems discussed by Mr Riehl. Since these observations have not yet been printed in extenso, even though some partial results have been published, I shall take the liberty of calling attention to two groups of deductions from the Mogadiscio data, both related to Mr Riehl's analyses.

1. The steadiness of the airflow in the "subcloud" or ground layer in the tropics can be seen quite nearly from the surface winds, whenever the relief is slight or non-existent, as in Mogadiscio, where the shore is quite flat. The observations were made with a Universal Anemograph of the Dines-Fuess type, at a height of 20 m. above sea level, and the records were carefully studied with respect to lateral turbulence. Fig. 1 (not published previously) represents in the lower half the hours of the Polar Year 1932—1933 with nearly laminar structure of the surface winds, i. e. quasi-uniform flow without lateral turbulence. The upper portion represents the hours with anomalous wind directions.

From this figure it is readily seen that there are *two* transition periods from one monsoon to

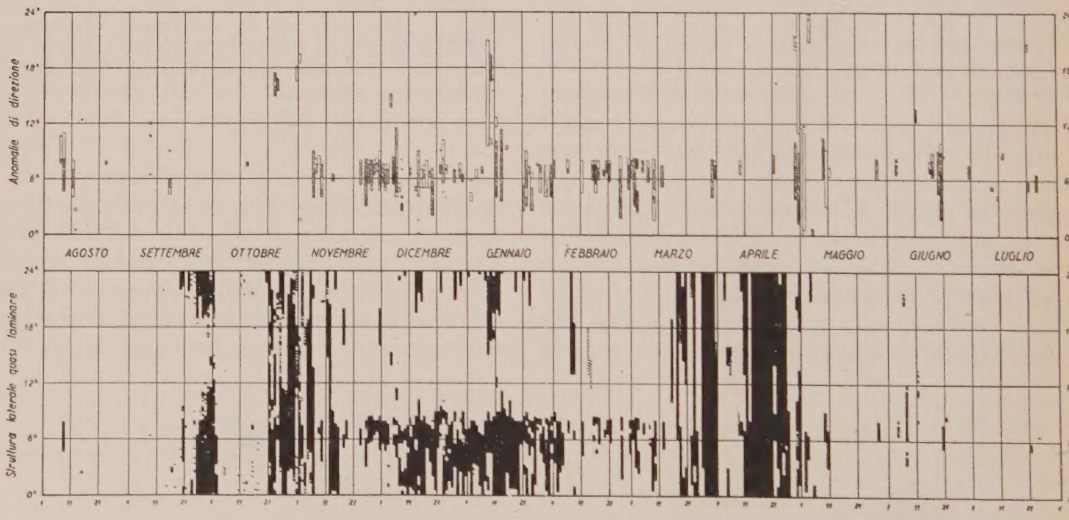

Fig. 1. Lower half: Dark areas or lines indicate hours with quasi-uniform flow in the surface layers (absence of lateral turbulence). Note that transition periods from winter to summer monsoon and vice versa have a double structure, suggesting in each case the passage of two Intertropical Convergence Zones.
Upper half: Shaded areas or vertical lines indicate hours with anomalous wind direction.
Both diagrams are based on surface wind observations (20 m above sea level) from Mogadiscio, Italian Somaliland, during the International Polar Year 1932—1933.

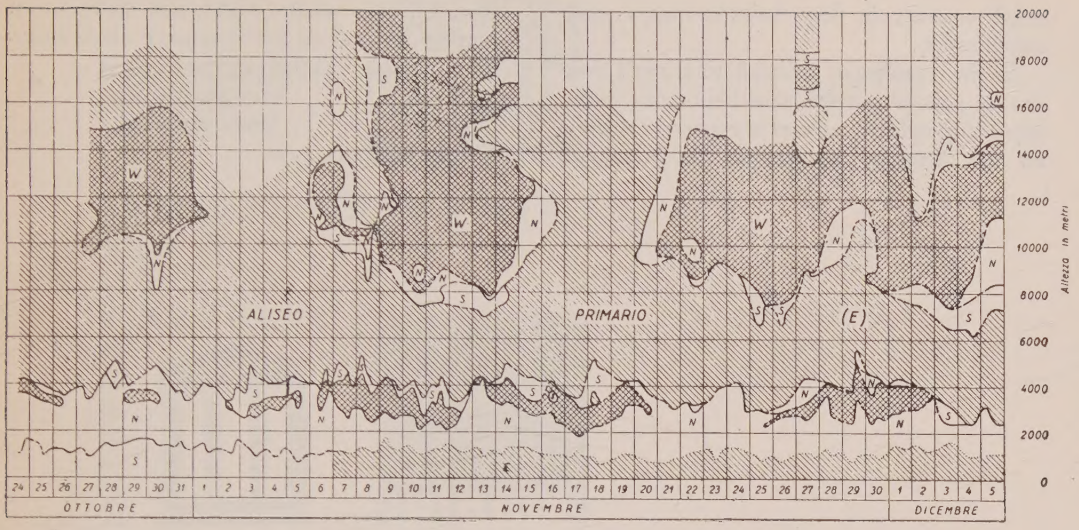

Fig. 2. Distribution of Basic Easterlies (Urpassat Aliseo Primario) and of other dominating upper winds during the period of maximum heights and maximum density of the pilot balloon observations made at Mogadiscio, October 24—December 5, 1932 (second part of the transition period).

the other, with the clear result that both transition periods (spring and fall) are characterized by *two* Intertropical Convergence Zone (ICZ) passages (see also the discussion by H. Flohn, 1949, Zeitschrift für Meteorologie, 3, pp 240—246), indicating a double structure of the ICZ, both in spring and in fall. During the transition periods the main wind direction is nearly the same as during the preceeding monsoon (the wind speed is less); hence, from this view point the principal ICZ is always the second one.

During the early morning hours, especially during the winter months of the Northern hemisphere (NE monsoon), the presence of a temperature inversion at low levels has the effect of giving the land breeze a quasi-uniform (laminar) structure. During the transition periods between the monsoons a similar inversion appears, again resulting quasi-uniform (laminar) flow, as shown in the lower portion of Fig. 1.

2. More than 1300 pilot-balloon ascents were made at Mogadiscio during the International Polar Year and from these data the mean picture of the tropospheric wind field has been deduced. The results were some time ago published (Zeit. f. Geophysik, 10, 1934, pp 360—368). However, it is only quite recently that some detailed inferences were made public (Geofisica pura e appl., XIV; 1949, pp 108—119), particularly with regard to the period October 24—December 5, 1932, this time interval being especially propitious for pilot balloon observations to great heights. The results obtained for this period show quite clearly, besides the evident steadiness of the ground layer and of the middle troposphere (Easterlies = Urpassat = Aliseo), the variability of the low part of the advection layer and particularly the fluctuations of the upper troposphere, where the easterlies intermittently are replaced by west winds. The time scale of the order of twelve to fifteen days is in accordance with Mr Riehl's indication "roughly two to three weeks". While the maximum speed observed in the upper troposphere easterlies was of the order of 50 m. p. s., at 14 km above sea level, in January, the greatest velocity observed in the westerlies was only about one half of this value or 26 m. p. s., at 15 km above sea level, at the end of November. Although the ascents in other months seldom reached as great heights as during this special period, it seems possible to infer that the upper troposphere westerlies are most frequent in the months from September to December, absent in March and rare in June, July, August (SW-Monsoon). Interesting is also the result that in Mogadiscio *no easterlies* are present in September, at least not below the tropopause level. This suggests that Mr Riehl's scheme for the

zonal currents in low latitudes needs some amplification.

In the 1949-paper referred to above I have already pointed out the great importance of the upper tropospheric currents in equatorial regions for the "steering" of weather also in middle latitudes. It is to be hoped that an increased number of stations for systematic observations soon may be organized in the tropics.

Very truly yours
Mario Bossolasco

Milano June 6th, 1950.

CORRIGENDUM

February issue of Tellus 1950, H. Riehl: "On the Role of the Tropies in the General Circulation of the Atmosphere."

The author has advised that unfortunately figs. 2 and 13 have been reversed and so have figs. 10 and 11.